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CHAPTER 3

PETROGRAPHY AND MINERALOGY

3.1 INTRODUCTION

The evolution of igneous rocks is in part recorded by mineral textures within the intrusives and partly by the mineral chemistry. Study of disequilibrium textures and mineral compositions has been an effective way of determining the role of magma mixing or mingling in silicic to intermediate rocks. The criteria for the identification of magma mingling aside from the occurrence of enclaves has been described by numerous authors (including Vernon 1983; Hibbard 1991; Vernon 1991a; Andersson & Eklund 1994; Wallace & Carmichael 1994; Gençalioglu Kuscu & Floyd 2001) and is summarized here into two categories.

(1) Textural or petrographic criteria: including disequilibrium textures such as sieved or dusty plagioclase, rounded and embayed crystals, reaction rims on minerals.

(2) Mineral chemistry criteria: including the occurrence of heterogeneous core to rim phenocryst compositions and normal and reverse mineral zoning.

The list within the categories is not intentionally extensive or exclusive.

The study of the textural development of igneous rocks can identify early stages of crystallization when crystal growth proceeds relatively uninhibited, followed by the linking up of crystals into a framework, and then by the interstitial crystallization of the residual magma. The presence of phenocrysts which enclose small crystals record a snapshot of the early-formed crystals in the magma. The primary mineral assemblage records the effects of pressure, temperature and water content. Compositional and textural features of phenocrysts in the mixed magmas can give clues about the composition and temperatures of the end member magmas (Wallace & Carmichael 1994).
In this chapter the mineralogy, textures and mineral chemistry of the felsic, mafic and intermediate rocks of the Sybella Batholith are presented and evaluated to see if features indicative of mixing are present and consistent with field relationships.

3.2 PETROGRAPHY

Modal compositions in terms of quartz, alkali feldspar and plagioclase are plotted in Figure 3.1 and listed in Appendix II. Figure 3.1 shows that the mineralogy of the mafic rocks is dominated by plagioclase, pyroxene, hornblende, biotite, apatite and opaque minerals and the granitic rocks are largely K-feldspar, quartz, plagioclase, biotite ± hornblende and titanite. The rocks identified as hybrids in outcrop have an intermediate composition relative to the mafic and felsic end-members, and commonly contain phenocrysts sourced from the two different end-member magmas. Biotite occurs in all rock types and is magmatic in origin and/or present as a replacement of amphibole. Pyroxene and hornblende occur in the dolerites and some of its hybrids. Apatite and opaque minerals are common accessory minerals in all rock types, while titanite and zircon occur in hybrids and granites.

The composition of minerals in these intrusions was determined using an electron probe microanalyzer. Each composition represents an average of several grains, each of which is the average of several analyses. Representative microprobe analyses of feldspar, amphibole, biotite and pyroxene of the mafic and felsic rocks are listed in Appendix III, and a summary of the textures and mineral compositions are given in Table 3.1 and Figure 3.2. Mineral classifications are given in Figures 3.3-3.7.

3.2.1 Mafic Intrusive Suite

Mafic intrusions that are represented throughout the development of the Batholith are divided into two groups: the Mosses Tank Dolerites and the Mafic Hybrid Complex as discussed in Chapter 2. The intrusive phases of the Mosses Tank Dolerites (in the north of the Batholith) have various textures and mineralogy including biotite and plagioclase porphyritic textures to equigranular fine- to coarse-grained textures. The dolerites are intruded by sheets of microgranite. Interaction between the different phases of dolerite
Figure 3.1 (A) Modal classification of geochemically analysed samples

Figure 3.1 (B) Modal mineralogy of the different rock types of the Sybella Batholith
<table>
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<th>Table 3.1 Summary of feldspar, biotite and amphibole compositions in different rock types, Syella Batholith.</th>
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<td><strong>Plagioclase</strong></td>
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**Notes:**
- An, Ab, Or, and Na+K refer to the anorthite, albite, ortorhochilite, and potassium content, respectively.
- Mg No. refers to the magnesium number.
- Na+K and Mg No. values are given in percentage.
- The table includes compositions from different rock types, including Main Phase Granite, Mafic hybrid, Microgranite, and Mosses Tank dolerite.
Figure 3.2  Mineral composition data illustrated on ternary and binary diagrams (a) Na\textsubscript{2}O - FeO - MgO plot highlights the Fe-rich composition of the biotite and amphibole minerals in the Main Phase granite; (b) Na\textsubscript{2}O+K\textsubscript{2}O - Al\textsubscript{2}O\textsubscript{3} - CaO plot of the Main Phase Granite Suite illustrates the intermediate nature of the interpreted hybrids. The K-feldspar composition shows tight clustering, including those xenocrysts from the Mafic Hybrid Complex; (c) Na\textsubscript{2}O+K\textsubscript{2}O - Al\textsubscript{2}O\textsubscript{3} - CaO plot of the Microgranite Suite highlights the amphibole alteration of pyroxenes in the Mosses Tank Dolerite; (d) Na\textsubscript{2}O+K\textsubscript{2}O - SiO\textsubscript{2} plot shows the intermediate nature of the interpreted hybrids in the Main Phase Granite Suit in Hbl and Pl compositions; (e) Na\textsubscript{2}O+K\textsubscript{2}O - SiO\textsubscript{2} plot shows the intermediate nature of the Pl in the interpreted hybrids.
Figure 3.3  Amphibole classification Leake et al (1997). (a) Calcic amphibole (Na+K) > 0.5 (b) Calcic amphibole (Na+K) < 0.5; (c) Fe$^3+$ Fe$^2+$ (calculated from Holland & Blundy, 1994) versus total Fe$^3+$/Fe$^2+$.
Figure 3.4  (a) Biotite classification;  (b) Mg/(Mg+Fe) versus total Al;  (c) Mg/(Mg+Fe) versus Si. Symbols as for Figure 3.2.
Figure 3.5  Pyroxene classification. Tie lines indicate co-existing pyroxenes in the diorite (sample 315) and in the dolerite (sample 18). The diorite pyroxenes are more Fe-rich which is consistent with other mineral compositions.
Figure 3.6 An content of plagioclase grains: core and rim compositions. Phenocrysts are indicated by tie lines between compositions. Reverse and normally zoned phenocrysts occur in mafic hybrid rocks.
Figure 3.7  Or content of K-feldspars; groundmass grains and phenocrysts.
is common, but with limited local hybridization with the microgranite. Dolerite and diorite to quartz-diorite intrusions of the Mafic Hybrid Complex (MHC) are pre- to syn-emplacement of the Main Phase granite. Compositional variation within the MHC is common due to inferred hybridization at depth and interaction at emplacement level where it was intruded by other rapakivi granitoids and Main Phase granites. Apart from the development of mylonites in local shear zones, all the mafic intrusions display igneous textures.

MOSSES TANK DOLERITES

The dolerite intrusions are fine- to coarse-grained with ophitic to sub-ophitic textures. They typically consist of plagioclase, clinopyroxene ± orthopyroxene, hornblende, biotite, magnetite/ilmenite with minor amounts of pyrite, chalcopyrite, quartz and apatite. Two distinctive groups of dolerites occur with one having a higher and the other a very low magnetic susceptibility respectively (Figure 2.4). Field relationships indicate that these two dolerite groups are coeval, with magnetite-rich enclaves pillowing within an ilmenite-rich dolerite (Figure 2.10).

The magnetite-rich dolerite is fine-grained and composed primarily of plagioclase laths and clinopyroxene ± orthopyroxene (with variable alteration to amphibole) and biotite (Figure 3.8). Plagioclase grains show a distinctive core (An₆₃₋₅₇) and rim (An₄₈₋₅₃) morphology and the core is variably retrogressed to sericite. Fe-rich orthopyroxene (Ferrosilite >Fs₅₀) is strongly fractured, irregularly shaped has fine-grained opaque minerals throughout and displays a lamellar structure (Figure 3.8ac). It is commonly surrounded by clinopyroxene (Diopside Wo₄₆₋₄₈ En₃₁₋₃₅ Fs₁₉₋₂₂) (Figure 3.5, 3.8c). Late-stage magmatic amphibole occurs as rims around pyroxene (Figure 3.8b) and post-magmatic amphibole alteration is associated with veins (actinolite to magnesiohornblende; Figure 3.8a). Biotite is commonly a phenocrystic mineral and occurs as individual large flakes (Figure 3.8b), as an interstitial mineral, and also has an acicular form (Figure 3.8a). These biotites are interpreted to be of magmatic origin. Biotite is commonly associated with coarse-grained opaque minerals. Biotite also occurs with amphibole with poorly defined or blurred grain boundaries suggesting an alteration origin (Figure 3.8a). Quartz is rare and occurs interstitially as one of the last minerals to crystallize. Apatite is blebby to acicular.
Figure 3.8  Mosses Tank Dolerite

(A) Subophitic textured dolerite with large plagioclase (with a dusty sericitised core and clear rim) and clinopyroxene grains (with ilmenite exsolution). Biotite occurs as large primary magmatic grains and as smaller secondary grains that has a more acicular mineral growth (indicated by arrow). Biotite also occurs with titanite as alteration around opaque minerals. Apatite is acicular to blebby. Hornblende occurs as alteration around pyroxene (sample 18); PPL.

(B) Large biotite flakes occur throughout the diorite. The biotite grains have inclusions of clinopyroxene rimmed by amphibole indicating changes within the magma chamber during crystallization. Larger clinopyroxene grains within the groundmass commonly have cores of orthopyroxene. Acicular apatite is common (sample 51); PPL.

(C) Strongly fractured and irregularly shaped orthopyroxene is surrounded by clinopyroxene. Coarse-grained opaque minerals within the groundmass are surrounded by biotite alteration. Apatite is rounded and blebby (sample 18); PPL.
The contact of magnetite-rich dolerite pillows and groundmass dolerite shows a mineral zonation with an increased proportion and alignment of plagioclase at the margin of the pillows (Figure 3.9a).

The ilmenite-rich dolerite is medium- to coarse-grained. Pyroxene occurs locally as relics in amphibole (Figure 3.9b). Clusters of subhedral to anhedral amphibole crystals with inclusions of tiny quartz blebs (Figure 3.9b) may be evidence that the amphibole formed through a reaction involving primary, near liquidus pyroxene:

\[ \text{pyx} + \text{melt} \rightarrow \text{hbl} + \text{qtz} \]  
(Bowens reaction series).

The amphibole also has an inclusion-free rim, possibly indicating an overgrowth after the replacement of pyroxene. Plagioclase commonly occurs as laths, but also as irregular zoned crystals in the groundmass. Biotite occurs as dark to medium red brown flakes and as large clusters that surround the oxides. The opaque minerals are skeletal or rods and are commonly mantled by titanite (Figure 3.9c). Acicular to blebbly apatite is interstitial to the plagioclase lath and hornblende framework, along with small elongate grains of biotite, amphibole and rare quartz (Figure 3.9c).

MAFIC HYBRID COMPLEX

The units of the Mafic Hybrid Complex range from fine- to coarse-grained and from diorite to quartz-diorite. The compositional contacts within the complex range from transitional to sharp and cuspate. These mafic rocks are separated into two groups based on their mineralogy.

The first group includes rocks with the most mafic composition, have ophitic to sub-ophitic textures and do not have xenocrysts of quartz and K-feldspar. They are primarily composed of clinopyroxene, orthopyroxene, plagioclase and biotite with minor amounts of hornblende, interstitial quartz, apatite, and opaque minerals (tr. pyrite) as accessory phases. Fe-rich orthopyroxene (ferrosilite >Fs60) occur as large elongated and cracked prisms with ilmenite exsolution lamellae, and fine-grained opaques (Figure 3.10a-c). Clinopyroxene (diopside En30-29 Wo47-48 Fs22-23 and augite) is greenish in colour and forms smaller, more irregular crystals, and commonly partly
surround the orthopyroxene (Figure 3.10ac). Plagioclase occur as large laths, forming a
subophitic texture and range from labradorite (An_{58-68}) core to andesine (An_{42}) rim.
Biotite occur as early-formed single coarse-grained flakes or clusters that are red-brown
in colour (Figure 3.10b). Small pyroxenes and plagioclase grains and coarse-grained
opaque minerals are commonly included in the biotite, along with some apatite crystals.
Amphibole is ferro-edenite (classification of Leake & et.al 1997; Figure 3.3), is late- to
post-magmatic, and formed at the expense of early pyroxene. It occurs as alteration
rims around pyroxene grains, and in the groundmass (Figure 3.10c). Apatite occurs as
large crystals and as smaller blebs. Quartz is interstitial. This group of mafic rocks is
interpreted as having been the least contaminated by felsic magmas.

The second group, although similar to the first, contains more abundant amphibole, K-
feldspar and quartz, the latter occurring as interstitial groundmass phases indicating late
crystallization. Clinopyroxene is irregular in shape, contains numerous fine-grained
opaque minerals and ilmenite exsolution and commonly surrounds orthopyroxene
grains. The clinopyroxene show partial to complete replacement by amphibole (ferro-
enite; Figure 3.11a). Plagioclase occur as phenocrysts and as prismatic grains in the
groundmass, and displays a variety of zoning patterns (Figure 3.11bc). Plagioclase
grains show core and mantle compositions of An_{45-49} and a normally zoned rim (An_{36}).
Reverse zoning occur in some phenocrysts with An_{49} core and rim An_{53} (Figure 3.6).
Other plagioclase phenocrysts show resorption in their zoning patterns (Figure 3.11c).
Some plagioclase phenocrysts have inclusion-rich cores, with pyroxene, biotite and
apatite, and an inclusion-free rim, similar to sieve textured volcanic plagioclases. The
inclusions commonly appear to be crystallographically controlled (Figure 3.12a) but
may also be irregular and confined to a particular zone within the phenocryst (Figure
3.12b). Similar to the more mafic rocks of the first group, biotite occur as large flakes
in the second group. However, in regions that are rich in quartz and K-feldspar, biotite
occur as a biotite-quartz symplectite. A secondary growth of acicular biotite (and
apatite) has also nucleated on the early large biotite grains, and radiates outwards
(Figure 3.12c). Apatite formed a variety of morphologies, from large skeletal crystals
with hollow cores to parallel and/or radiating arrays of acicular apatite needles in the
interstitial minerals of quartz and K-feldspar. Quartz and K-feldspar (Or_{90}) form as
interstitial minerals and as grains within the groundmass. In the more quartz-rich
Chapter 3
Petrography and Mineralogy

A

B

C

3-15
Figure 3.9  Mosses Tank Dolerite

(A) Contact of medium-grained dolerite pillow within the finer-grained host dolerite. The margin of the dolerite pillow has numerous plagioclase laths that are aligned to the contact with the host. Composition of both the pillowed and host dolerite is similar although grain size varies (sample 19) XPL.

(B) Relic clinopyroxene occur within amphibole grains. Quartz inclusions in amphibole clusters are very common and result from the replacement of clinopyroxene. Biotite (and titanite) occur as alteration around clustered rods of ilmenite (sample 17); PPL.

(C) Late stage minerals grow in cavities developed in the plagioclase-amphibole framework. The minerals include late-stage quartz, and a more acicular form of hornblende and apatite that grew inwards from the edge of plagioclase crystals (sample 17); PPL.
Figure 3.10

(A) Orthopyroxene with ilmenite exsolution partly surrounded by clinopyroxene in a groundmass of predominantly plagioclase, biotite, clinopyroxene (CPL: sample 202);

(B) Lamellae occur within the Fe-rich orthopyroxenes. Exsolution textures indicate that some of these pyroxenes are inverted pigeonites (PPL: sample 332);

(C) Large irregular orthopyroxene grains surrounded by clinopyroxene.

Clinopyroxene within the groundmass has minor alteration to amphibole.

Clinopyroxene and coarse-grained opaques are included in large biotite flakes (PPL: sample 332).
Figure 3.11

(A) Amphibole alteration of pyroxene in quartz-diorite (PPL: sample 224);

(B) Plagioclase phenocryst with an inclusion-rich core (a), an inclusion free rim (b), and overgrowth (c). Inclusions are biotite, amphibole and pyroxene. Amphibole has partly altered clinopyroxene and also occurs as small grains in the groundmass (CPL: sample 224);

(C) Mineral zonation in a plagioclase phenocryst is truncated, which indicates it was not in equilibrium with the melt after its inclusion in the mafic hybrid magma. The thin inclusion-rich plagioclase overgrowth crystallised in the mafic hybrid (CPL: sample 410).
Figure 3.12

(A) Crystallographically aligned biotite and pyroxene inclusions in a plagioclase phenocryst (black arrows). The inner core has smaller and fewer inclusions than the outer core and the overgrowth has no inclusions (sample 202); PPL.

(B) Inclusion-rich zone in plagioclase phenocryst within a hybrid groundmass (sample 325); PPL.

(C) Radiating biotite (indicated by black arrows) is a late stage mineral growth in addition to the early phenocrystic biotite (not shown in this image) (sample 223); PPL.
Figure 3.13

(A) Radial mineral growth (amphibole and plagioclase) in a mafic enclave indicate rapid cooling and crystallization. Plagioclase phenocrysts have a sericitised core (sample 207); PPL.

(B) Apatite is acicular within a mafic enclave and commonly have “step like” patterns (sample 322); PPL.

(C) Radiating and step-like textures of acicular apatite in hybrid rock is indicative of rapid cooling and crystallization possibly resulting from mixing a mafic magma with a more felsic magma (sample 516); PPL.
diorites, intergrowths of feldspar and quartz occur at grain boundaries.

MAFIC ENCLAVES

Mafic enclaves occur within the Main Phase granites and the microgranites. In the undeformed phases of the granite, the mafic enclaves illustrate a magmatic texture with radiating mineral growth (Figure 3.13a). Spectacular patterns of acicular and hollow needles of apatite occur within the late stages of mineral growth (Figure 3.13bc).

3.2.2 Main Phase Potassic Granitoid Suite

The Main Phase granites can be broadly separated into two groups: K-feldspar megacrystic syenogranite and porphyritic syenogranite. The megacrystic syenogranite is most commonly recognized as the Main Phase granite of the Sybella Batholith. It contains large K-feldspar megacrysts and is dominantly biotite bearing, but has a small percentage of amphibole. The porphyritic syenogranite is a medium- to fine-grained granite, has minor biotite, high quartz content, K-feldspar megacrysts, and occur as late-stage small intrusions and dykes. The Main Phase granite suite illustrates a variety of deformation ranging from undeformed to mylonites. Deformation within the units is described and discussed in Chapter 5.

“MAIN PHASE” MEGACRYSTIC SYENOGRAINITE

The Main Phase biotite syenogranite is coarse-grained, heterogeneous with 1 to 5 cm K-feldspar megacrysts. The intrusions are typically composed of both perthite and microcline megacrysts, plagioclase, quartz, hornblende, biotite, titanite, accessory magnetite, apatite, zircon, allanite and fluorite. Phenocrystic minerals in the Main Phase granite include the K-feldspar megacrysts, plagioclase and quartz (Figure 3.14).

The K-feldspar megacrysts generally occur as single grains, however in areas that are strongly deformed, the megacrysts are partially to completely recrystallized to clusters of microcline (with triple point junctions) and occur as augen (See Chapter 5, Figure 5.16a). The K-feldspar megacrysts are zoned, commonly with a euhedral to anhedral core (Or86) and an overgrowth or rim (Or94) that extends into the surrounding groundmass (Figure 3.14-15). The core is generally inclusion free, but may contain irregular quartz growths and small plagioclase crystals. The inner to outer core boundary can be defined by a compositional change in the feldspar and also by the
Figure 3.14  Main Phase Granite

Large altered plagioclase phenocrysts, perthitic K-feldspar phenocrysts, large rounded quartz grains are characteristic of the Main Phase granite. Plagioclase phenocrysts have a sericitised core. K-feldspar is zoned. Biotite, amphibole quartz and feldspar occur within the groundmass. Titanite and opaques (magnetite) are common (Sample 219); XPL.
Figure 3.15  Main Phase Granite

(A) Edge of K-feldspar megacryst with an individual plagioclase grain included in the rim as part of a mantle and zone of quartz inclusions (arrow). Plagioclase phenocryst (bottom tight) is also zoned (sample 242); XPL.

(B) K-feldspar megacryst with plagioclase overgrowth forming the rapakivi texture (sample 351); XPL.

(C) The occurrence of myrmekite (arrow) at the margins of K-feldspar megacrysts may be deformation induced or result from late stage water saturated crystallization (sample 226); XPL.
presence of mineral inclusions. The outer core commonly contains small plagioclase, quartz, biotite, opaque minerals, apatite, titanite and rarely amphibole. A rim or mantle may occur with large individual plagioclase phenocrysts (An_{32-24}) being incorporated into the outer rim zone (Figure 3.15a) or as an overgrowth crystallographically aligned with the K-feldspar megacryst which formed a partial rapakivi texture (Figure 3.15b). Plagioclase phenocrysts commonly have a sericitised core and unaltered rim. Quartz phenocrysts are rounded and in many rocks they have been recrystallized to quartz clusters (Figure 3.14). In the strongly deformed Main Phase granites, the quartz phenocrysts occur as quartz ribbons.

The groundmass is coarse-grained and contains abundant K-feldspar (Or_{92-93}), quartz, plagioclase (An_{20-22}), biotite and hornblende as well as the accessory minerals. Amphiboles are hastingsite and have inclusions of biotite, plagioclase and opaque minerals. Biotite is annite in composition and may be altered to chlorite. Titanite is anhedral (rarely euhedral) and occurs in the groundmass, as inclusions in biotite and in some places associated with opaque minerals. Apatite is bleby, zircon is commonly zoned, and when included in biotite they have radioactive halos. Fluorite occurs interstitially. Myrmekite lobes (quartz and plagioclase intergrowths) occur along the margins of the K-feldspar megacrysts, and within the megacrysts at the boundaries with plagioclase inclusions (Figure 3.15c). Myrmekite is common within the groundmass.

PORPHYRITIC AND APLITIC SYENOGRANITE
The porphyritic syenogranite contains phenocrysts in an even, fine- to medium-grained groundmass. K-feldspar, quartz and plagioclase phenocrysts are similar to those described in the megacrystic Main Phase granite (Figure 3.16ab). The groundmass consists of K-feldspar, plagioclase and quartz, biotite (chlorite alteration), accessory zoned zircon, titanite and rare fluorite. Groundmass textures vary with some pegmatitic intergrowths of quartz and K-feldspar (Figure 3.16c). Myrmekite and recrystallization of the groundmass minerals and quartz phenocrysts is common in the deformed granite.
Figure 3.16 Porphyritic Granite

(A) Quartz phenocrysts in K-feldspar porphyritic granite. The quartz shows chessboard extinction indicative of high temperature submagmatic deformation (sample 324); XPL.

(B) Zoned plagioclase phenocryst in porphyritic granite. Zonation (dotted line) has been truncated (arrow) and has been overgrown with an inclusion rich rim (sample 324); XPL.

(C) Pegmatitic textures in the matrix of the K-feldspar porphyritic granite (sample 324) XPL.
3.2.3 Microgranite Suite
There are two texturally different granites within the microgranite suite, with the dominant (and younger) phase being a microgranite. The oldest phase is medium- to coarse-grained and in some instances has small euhedral K-feldspar phenocrysts. The main body of the microgranite phase is medium- to fine-grained and equigranular. The microgranites are characterized by a high K-feldspar and quartz content and low mafic mineral content. Compositionally syenogranites, they are composed of K-feldspar, quartz and plagioclase, with minor biotite, and accessory magnetite, zircon, titanite, apatite and fluorite (Figure 3.17). The plagioclase is strongly sericitised and commonly replaced by secondary muscovite (Figure 3.17a). Alkali feldspar is microcline, perthitic and euhedral in shape (Figure 3.17a). Biotite is partly to completely replaced by chlorite. Quartz is interstitial. Titanite occur as rims around opaque minerals. Accessory apatite is bleby and fluorite is interstitial (Figure 3.17b).

3.2.4 Mafic-Felsic Hybrids
Hybrids are formed from mixing Main Phase granite and magmas from the Mafic Hybrid Complex, and also microgranite and Mosses Tank Dolerite. As described in Chapter 2, much of the mixing with the Main Phase granites occurred prior to emplacement. This is exemplified by rapakivi granitoid intrusions, and the xenocrystic units of the Mafic Hybrid Complex (Section 3.2.1). In some instances however, there are gradational boundaries between the hybrid or granite phases and the diorite, which indicates local mixing related to emplacement. The microgranite hybrids formed during emplacement with localized mixing. These hybrids have a very similar phenocrystic mineral assemblage and groundmass mineral distribution to that of the Main Phase granite and dolerite hybrids. The hybrids are described in terms of phenocrystic and xenocrystic mineralogy and also groundmass compositions and textures.

Hybrids display mineral assemblages from both end-member magma compositions: the Mafic Hybrid Suite and the Main Phase granite or the Mosses Tank Dolerite and microgranite. The hybrid group of rocks is variable in composition and appearance. The hybrids most similar to the mafic end-member have a broadly similar mineralogy and texture to the mafic rocks, but with K-feldspar, plagioclase and quartz xenocrysts. This is transitional to a hybrid with similar mineralogy to the felsic end-member (Main
Figure 3.17 Microgranite

(A) Altered plagioclase (muscovite flakes) with a rim (black arrow) occurs with perthitic euhedral K-feldspar as larger grains in a finer grained and commonly recrystallized groundmass. White arrow (inclusion rich zone) may indicate the extent of the original K-feldspar grain, which has been overgrown. The granite has a low mafic mineral content (sample 131); XPL.

(B) Muscovite occurs as alteration of plagioclase and rarely in the matrix. Biotite occurs within the groundmass although it is commonly partially altered to chlorite (sample 291) XPL.

(C) Plagioclase grain is included in rim zone of K-feldspar. Plagioclase has a sericitized core and unaltered rim (sample 291) XPL.
Phase granite or microgranite), but with elevated mafic component that may be heterogeneously distributed in clusters, and minerals exhibiting disequilibrium textures. These disequilibrium textures include but are not limited to: sieved or dusty plagioclase, rounded and embayed crystals, and reaction rims on minerals.

**Plagioclase phenocrysts**

Plagioclase phenocrysts are euhedral to rounded in shape and commonly have several stages of growth. An inclusion-free core may be present which was overgrown by an inclusion-rich “outer” core (Figure 3.18). This core sometimes shows compositional zonation that is truncated by a dissolution surface, reflecting a change in melt equilibrium. This was generally followed by another stage of growth such as an inclusion-free rim zone (or overgrowth) similar in composition to the plagioclase grains in the groundmass (Figure 3.18-19). The core is commonly partially sericitised, and the rim unaltered. In many of the mafic units such as the mafic hybrid complex or mafic enclaves, the inclusion-rich core or outer core formed a spongy texture (Hibbard 1991) as discussed in Section 3.3.

Compositional zoning was observed both petrographically and in backscatter microprobe images (although the resolution was not as good in the latter; Figures 3.19-20). Phenocrysts may be normally zoned or reversely zoned (Figure 3.6). In the Main Phase granite, the phenocrysts have normal zonation, although initial transects from core to rim indicate variation in this general pattern. Reverse zoned plagioclase phenocrysts occur in the hybrid diorites.

**K-feldspar xenocrysts/phenocrysts**

K-feldspar phenocrysts exhibit complex zoning in the form of inner core (± outer core), rim and (±) overgrowth that extends into the groundmass. The cores of the K-feldspar phenocrysts in the hybrids are similar to those in the granites. The core may be euhedral to anhedral, commonly inclusion free, but may contain irregular quartz growths and small plagioclase grains. The rim zone commonly contains small inclusions of plagioclase, quartz, biotite, opaque minerals, apatite, titanite and rare amphibole (Figure 3.21).
K-feldspar phenocrysts commonly have an outer zone of plagioclase grains, forming a rapakivi texture. This texture may comprise a single grain of plagioclase that has overgrown a core of K-feldspar (Figure 3.21a), as multiple plagioclase grains that are included in or partly surrounded by the K-feldspar (Figure 3.15a, 3.21b), as a complex growth of quartz, K-feldspar and plagioclase (Figure 3.21c), or as a combination of the three.

Some phenocrysts are characterized by irregular intergrowths of plagioclase and K-feldspar (Figure 3.22). In the mafic hybrids the K-feldspar phenocrysts rims have a lower Or content than the core, defining a reverse zonation (Figure 3.7). Microprobe transects from core to rim in K-feldspar phenocrysts show considerable variation in compositions with Or contents ranging from 60 to 94 (Figure 3.7).

**Quartz xenocrysts**

Quartz xenocrysts occur in all the hybrids, where they are rimmed by clinopyroxene ± amphibole (in the more mafic compositions) or by amphibole (in the more silica rich granite hybrids) forming ocelli texture (Figure 3.23a and b). The quartz is typically rounded and may have irregular embayments. Some of the smaller mafic minerals appear fully enclosed in the quartz, although most commonly extend into the surrounding groundmass.

**Ground mass mineralogy and textures**

In the Mafic Hybrid Complex, the distribution of minerals in the groundmass is relatively homogeneous, and aside from the phenocrystic minerals assemblage and the predominance of amphibole as the common mafic mineral, these hybrids have late stage acicular apatite and biotite, and skeletal zircon grains (Figure 3.23c). K-feldspar and quartz occur interstitially and granophyric intergrowths of quartz and feldspar are found within the groundmass. In the more silica-rich hybrids, such as the rapakivi granitoids, the mafic minerals (biotite and amphibole ferro-edenite to hastingsite) are distributed in clusters with plagioclase that resemble a diorite composition (Figure 3.23b). Many of the hybrids have very low Fe-Ti oxide contents, but have a very high proportion of titanite (Figure 3.1b).
Figure 3.18  Plagioclase zonation

(A) Zoned plagioclase phenocryst in hybrid with pyroxene inclusions (sample 325); XPL. (B) Zoned plagioclase phenocryst in mafic hybrid. Plagioclase has an inclusion rich inner core (a), zones (b) and (c) are outer core which are relatively inclusion free, and (d) is a rim zone that is an overgrowth which extends and includes minerals (pyroxene, biotite) in the matrix (sample 325); XPL. (C) Multiply zoned inclusion-rich plagioclase phenocryst in hybrid. Inner core is inclusion free whilst the outer zones have numerous quartz inclusions that are box-shaped or edges that form 90 degree angles, implying some crystallographic control (sample 410); XPL.
Figure 3.19
Backscatter image from the microprobe shows a zoned plagioclase phenocryst in a hybridised mafic rock. The zonation is also marked by a zone of inclusions and compositional change shown by variation in colour.

The enlarged area of the phenocryst shows weak compositional zoning (arrow) and also inclusion-rich overgrowth that extends into the matrix.

A photomicrograph of an area within the phenocryst shows the optical zonation.

Sample 410, Mafic Hybrid Complex.
Figure 3.20

(A) Backscatter image (left) and photomicrograph (right) of zoned plagioclase phenocryst with an inclusion rich core. It is included in the rim zone of a K-feldspar phenocryst in mafic hybrid (sample 410).

(B) Backscatter image (left) and photomicrograph (right) of inclusion-rich plagioclase phenocryst in rapakivi hybrid. Inclusions are biotite (elongate) and amphibole (rounded) (sample 314).
Figure 3.21 Rapakivi texture

(A) K-feldspar phenocryst in medium-grained hybridized microgranite completely surrounded by a single plagioclase grain forming a mantle (sample 281); XPL.

(B) K-feldspar phenocryst in medium-grained microgranite hybrid with numerous plagioclase grains in the mantle forming the rapakivi texture (sample 281); XPL.

(C) The rapakivi hybrid is characterized by large K-feldspar phenocryst core with a relatively wide outer zone of intergrown plagioclase, K-feldspar, quartz and mafic minerals in hybrid (sample 314); XPL.
Figure 3.22

Backscatter images and photomicrograph of a predominantly K-feldspar phenocryst with irregularly intergrown plagioclase and minor quartz in a mafic hybrid. The large inset shows the boxy nature of the intergrowth indicating some crystallographic control on the minerals (sample 410).
Figure 3.23  Hybrid textures
(A) Clinopyroxene and amphibole grains rimming quartz xenocryst in hybrid (sample 334); PPL.
(B) Quartz xenocryst with grains of amphibole forming a rim. A cluster of amphibole grains also occurs within the hybrid (sample 314); PPL.
(C) Skeletal zircon occurs within the mafic hybrids (sample 332); XPL.
(D) Backscatter image of resorbed and zoned zircon in mafic hybrid (sample 410).
3.3 DISCUSSION OF MINERALOGY AND TEXTURES

3.3.1 Mafic Mineral Compositions

Mafic minerals are good indicators of intensive variables, in particular temperature and the water and oxygen fugacities. The occurrence of magnetite and ilmenite relate to the relatively oxidized or reduced status of the solidifying magma, respectively. An example of the influence of water activity, and the application of textural criteria, is the reversal of the order of crystallization of augite, hornblende and biotite according to the Bowen Reaction series. In accordance with this theory, with increasing fractionation and water contents, hornblende is stabilized before biotite in calc-alkaline melts. However, Wones & Gilbert (1982) indicate that biotite may appear directly after augite and before hornblende (which is observed in the mafic units) if the magma is undersaturated with water. Biotite may also be a response to mixing a mafic magma with high-K felsic magma (Main Phase granite or similar magma) at depth.

The amphiboles are iron-rich, similar to those occurring in typical anorogenic granites as defined by Anderson & Bender (1989). The amphiboles are members of the calcic group and compositions range from hastingsite in the Main Phase granites to ferro-edenite in the hybrids, diorites and dolerites (Figure 3.3a). The compositional variation generally complements that of the coexisting biotite, although with higher Fe$^T$/Fe$^T$+Mg ratios (Table 3.1, Figure 3.2). However, post-magmatic amphiboles occur as rims on pyroxenes in the diorites (ferro-hornblende) and in the dolerites (magnesio-hornblende to actinolite; Figure 3.3b). These late stage amphiboles have significantly lower Na, K and Fe$^T$ contents, than those interpreted as magmatic. The Main Phase granites show higher Fe$^T$/Fe$^T$+Mg than either the dolerites or hybrids. Calculated Fe$^{3+}$/Fe$^{2+}$ ratios (after the method of Holland & Blundy 1994) are generally low, however the amphiboles in the hybrid microgranite are slightly higher (Figure 3.3c). The amphiboles are more Mg-rich in the diorites and hybrids than the granite and may reflect crystallization from magma with an increased oxygen fugacity related to the mixing. The initial mafic magma was undersaturated in water (because of early biotite—see previous), but when it mixed with relatively more hydrous granite, amphibole crystallized forming rims and also replaced pyroxene.
Biotite compositions vary significantly in Mg/(Mg+FeT) ratio (Table 3.1). The most iron-rich biotite occurs in the Main Phase megacrystic syenogranite with Mg/(Mg+FeT) ratio between 0.25 and 0.29 (Figure 3.2-3). Biotite grains from the other units have ratios that range between 0.37 and 0.49. The FeT/Mg ratio of biotite coexisting with K-feldspar and magnetite varies as a function of T, fO2 and fH2O (Wones 1981). High fO2 and low fH2O increases the Mg/(Mg+FeT) ratio of biotite. The most Mg-rich composition occurs in the microgranite hybrids, Main Phase hybrids, and rapakivi granitoids and may be a result of the increased oxygen fugacity during mixing. There is a small range in total aluminium content with the higher values found in the Main Phase granite and microgranite (2.55-2.76 Al p.f.u.), and lower values (2.33-2.55 Al p.f.u.) in the mafic units (Figure 3.4b). This is also reflected in the Ti composition, with the granites having lower Ti and the mafic and hybrids containing higher Ti.

Both orthopyroxene and clinopyroxene occur in the Mafic Hybrid Complex and Mosses Tank Dolerites however textures indicate that they are not in equilibrium with each other (Figure 3.5). The elongate grains of orthopyroxene (ferrosilite) have ilmenite exsolution lamellae and are commonly resorbed and cracked. Some of the lamellae lie in the (100) plane indicating that some of these are inverted pigeonite. The high Fe-content may indicate that the pyroxenes crystallized from a relatively fractionated mafic magma. Clinopyroxene is diopside in composition with some augite, and crystallized after the orthopyroxene.

The ilmenite lamellae in the pyroxenes and ilmenite within the early biotite suggest that the mafic magma was initially relatively low in fO2. Mixing of this reduced mafic magma with a more oxidized magma would lead to the formation of magnetite. This process would explain the occurrence of magnetite in the matrix and ilmenite in the early-formed minerals in the Mafic Hybrid Complex and rapakivi granitoids. Magnetite, the only oxide phase within the Main Phase granite, is commonly euhedral and occurs within the K-feldspar and groundmass. The biotite-titanite±hornblende mineralogy of the granites, the late crystallization of the hydrous phases and the euhedral Fe-Ti oxide mineralogy constitutes a magnetite series, implies low water content and an elevated oxygen fugacity (Wones 1989). The presence of late stage anhedral titanite in the hybrids results from oxidation related to mixing the Main Phase
3.3.2 Textural and Compositional Evidence for Mixing
Microstructural evidence consistent with the process of hybridization includes mantled quartz xenocrysts (“ocelli”), K-feldspar megacrysts with or without plagioclase mantles (“rapakivi”), and zoning discontinuities in plagioclase (Figures 3.11-23). Phenocrysts that were stable in the granitic magma, were partially resorbed in the mafic hybrids, and that produced rounded quartz and K-feldspar xenocrysts. New minerals then crystallized around the partly resorbed phenocrysts. The destabilized felsic phenocrysts that had mechanically been introduced into the more mafic hybrid are therefore considered as xenocrysts in their new environment (e.g. strong reverse zoning occurring at the rims of plagioclase phenocrysts; Figure 3.6). The quartz-diorites in the Mafic Hybrid Complex and rapakivi granitoids display the above evidence for disequilibrium. Clusters of mafic minerals and plagioclase within the rapakivi granitoids are interpreted to be relics from the incomplete mixing of the original magmas (Figure 3.23). Mineralogy of the hybrids, particularly the Fe-Ti oxides, mafic mineral compositions (mentioned in previous section) and undercooling generated acicular apatite, also reflects changes in magmatic $fO_2$, $fH_2O$ and temperature conditions.

QUARTZ OCELLI
Quartz “ocelli” are common in hybrid magmas and microgranitoid enclaves. Generally, they have a thin rim consisting of the early-formed mafic minerals (Vernon 1991a). These minerals include clinopyroxene ± amphibole ± biotite depending on the composition of the mafic melt. This texture is formed due to the instability of quartz grains (from a felsic magma) being incorporated by mixing into the mafic magma. Resorption of the quartz xenocryst takes heat from the immediately adjacent melt causing local undercooling, and the generation of numerous small mafic crystals (Castro et al. 1990; Vernon 1991a). A second phase of quartz grew on the original surface, and formed a rim that locked in the small mafic minerals within the rim zone.

K-FELDSPAR MEGACRYST
K-feldspar megacrysts are common in the hybrid rocks and in the megacrystic granites, which suggests a similar factor in their origin. The megacrysts in both the rapakivi granitoids and the granites are microstructurally very similar in their shape,
composition, zoning patterns, and inclusions (± mantle development), so it is likely that they grew in very similar environments, thus it is logical to interpret the megacrysts in the same way. There are two schools of thought regarding the origin of K-feldspar in granitic rocks. K-feldspar megacrysts have been interpreted as representing phenocrysts that grew from a melt (Vernon 1986; Hibbard 1965) or subsolidus K-metasomatism and subsequent porphyroblast growth or replacement of plagioclase phenocryst (Collins 1997; and discussed in Vernon 1986). The widespread uniformity of composition and textures of the megacrysts and general high-potassic composition of the suite of rocks supports the magmatic origin. The batholith does not show zones of high- and low-K composition as would be expected if it were due to metasomatic alteration and porphyroblast growth. Evidence of crystallization from a melt includes euhedral shape of the megacryst, magmatic alignment, conformity of size, zoning marked by inclusions and chemistry, and the euhedral shape and zoning of plagioclase inclusions. Potassium feldspar megacrysts represent multistage growth of a phenocryst from a melt, ending with some extension of growth in the near solid. The internal zones of tiny inclusions of plagioclase, biotite and quartz represent still stands in the growth of the host. The interpretation of the K-feldspar megacrysts representing mineral growth in a magmatic state is preferred for the Sybella Batholith.

RAPAKIVI TEXTURES

Rapakivi textures or mantles of plagioclase formed on magmatic K-feldspar occur in both the main phase granites and in the mafic and granitic hybrids. Theories concerning the origin of rapakivi texture fall into three groups – those dominated by magmatic processes, those involving subsolidus processes, and those invoking a combination of magmatic and subsolvus processes (Dempster et al. 1994). Some magmatic models argue that the texture results from a magmatic process and unrelated magma composition (Nekvasil 1991), whilst others suggest that the texture relates to the chemistry of the magmas. These models require a change in conditions to firstly allow K-feldspar to crystallize followed by a period of plagioclase crystallization that forms the mantle (e.g. degassing, decompression, and magma mixing; Cherry & Trembath 1978; Hibbard 1981; Bussy 1990; Nekvasil 1991). Early crystallization of K-feldspar from the melt will occur in K-rich magmas. Alternatively, subsolidus exsolution of K-feldspar and redistribution of albite in fluorine-rich rich magmas (Dempster et al. 1994)
may contribute to the formation of rapakivi texture.

The rapakivi textures within the Sybella Batholith intrusives varies significantly with a wide range of shapes, compositions, mineral zonation and intergrowths as seen in Figures 3.14-15, 22-23. The shape of the rapakivi feldspars has been variously attributed to mechanical abrasion, resorption (Nekvasil 1991), corrosion, exsolution and redistribution of plagioclase from the K-feldspar (Dempster et al. 1994), however the ovoid or embayed shape is likely to be a combination of the above processes. The initial concentric zonation of the K-feldspar core is attributed to magmatic processes, followed by overgrowth of mantle (or replacement of K-feldspar by plagioclase), although resorption of the core is evident in the embayed shapes (e.g. Figure 2.9). The rapakivi-textured phenocrysts may have a core of perthitic K-feldspar and a rim zone or mantle of varying textures and compositions (described in Section 3.2.4). The concentrically zoned K-feldspar cores contain quartz, plagioclase, mafic mineral and oxide inclusions. The quartz is irregular in shape but the plagioclase inclusions are commonly zoned and are euhedral to subhedral in shape, which indicates a magmatic origin. The mantling textures vary significantly, ranging from individual plagioclase grains attached to the surface of the K-feldspar grain, to a single large overgrowth of plagioclase, to intergrown crystallographically aligned K-feldspar and plagioclase that has irregular interstitial quartz, and plagioclase overgrowths that extend into the groundmass. The mantle may also include zones rich in mafic minerals. In some of the more mafic hybrids, some phenocrysts dragged along a felsic mesostasis rim after its capture from the granitic melt and that resulted in a trail of minerals extending from the phenocryst into the mafic groundmass. Hence, some margins of the rapakivi may be interpreted as granitic melt included with the K-feldspar, rather than having formed as an overgrowth whilst in the more mafic magma (Bussell 1991). Replacement of K-feldspar by plagioclase that resulted in a mantle of K-feldspar, quartz and plagioclase is not excluded as a process, as some partially mantled phenocrysts exhibit textures that could be interpreted as a result of this reaction. Because calcic plagioclase requires less silica in its lattice than in K-feldspar or sodic plagioclase, silica is left over to form quartz.
PLAGIOCLASE PHENOCRYSTSS

Plagioclase phenocrysts in Mafic Hybrid Complex and rapakivi granitoids show complex zoning patterns and discontinuities or truncation of mineral zonation that may reflect changes in composition through the mixing of magmas, possibly during the recharge of magma chambers (Hibbard 1981; Vernon 1991a). Reverse zonation in plagioclase phenocrysts in the mafic hybrids is common. The higher An content of the plagioclase rims represents growth whilst in the mafic magma, whereas the relatively more sodic cores represent a xenocrystic origin. Normally-zoned plagioclase phenocrysts in the rapakivi granitoids exhibit abrupt changes in mineral composition from a calcic core (grown whilst in a mafic magma) to a sodic rim (grown whilst in the more felsic hybrid magma).

Cellular plagioclase phenocrysts or inclusion-rich phenocrysts occur within the hybridized mafic units and formed as a result of incorporation of plagioclase into the mafic magma from the felsic magma. The plagioclase (andesine) became unstable as it heated in the hotter mafic magma and began to dissolve, forming the cellular zones. A cellular mantle is developed at conditions below the plagioclase liquidus in plagioclase crystals that are more sodic than those in equilibrium with the melt (Andersson & Eklund 1994). Above the liquidus the xenocryst suffered dissolution or melting and became rounded. Dissolution preferentially follows imperfections in the crystal, such as cleavage planes (e.g. Figure 3.11b, 12, 18), which caused the cellular texture. The spongy xenocrysts are penetrated by the enclosing magma and droplets are trapped within the cells (Andersson & Eklund 1994). Pyroxene, amphibole, biotite, oxides and apatite crystallize within these regions and are commonly optically continuous. Inclusion-rich zones in the K-feldspar xenocrysts are thought to have resulted from similar high temperature dissolution and crystallization processes.

INTERGROWN PLAGIOCLASE-K-FELDSPAR XENOCRYSTS

The “motheaten”, irregularly intergrown xenocrysts in Figure 3.22 are predominantly K-feldspar in composition and occur within hybridized mafic diorites. The K-feldspar has small box-shaped or elongated patches of plagioclase throughout the mineral with larger patches of plagioclase crystal structure intergrown with quartz towards the center of the grain. The plagioclase is aligned parallel to former crystal faces. The outer zone
of the xenocryst has numerous small mafic mineral inclusions similar to that in the cellular plagioclase phenocrysts. The reaction of K-feldspar to plagioclase and quartz may result from dissolution and recrystallization after the incorporation of K-feldspar phenocrysts into the mafic magma from the felsic magma. After inclusion of felsic phenocrysts, the temperature was above the liquidus so the xenocryst suffered dissolution preferentially along imperfections in the crystal, such as cleavage planes (Andersson & Eklund 1994; Dempster et al. 1994). As dissolution continued, a cellular structure developed in the rim zone of the K-feldspar xenocrysts (similar to the inclusion-rich plagioclase phenocrysts). The enclosing magma penetrated the xenocryst and droplets were trapped within the cells forming rounded inclusions.

**ACICULAR APATITE**

Acicular apatite has been identified as a mixing texture (Hibbard 1991). Rapid growth of apatite in an internally quenched mixing system results in acicular rather than stubby apatite that was identified in the mafic and granite end members. Apatite forms spectacular growth patterns in arrays of radiating and subparallel groups of minerals in the interstitial phases.

### 3.3.3 Other textures

Quartz intergrowths are common in all rock types. They occur with biotite, amphibole and feldspars in the late stage minerals in the groundmass and as overgrowths on early-formed minerals resulting from the simultaneous crystallization of two minerals. Biotite phenocrysts in the mafic units have quartz intergrowths at the margins of the crystals possibly due to rapid crystal growth during chilling (resulting from a mixing event). Quartz and feldspar intergrowths in the groundmass of the porphyritic syenogranite dykes (Figure 3.16c) are a distinctive feature indicating rapid crystal growth due to cooling.

Myrmekite occurs both in the deformed and in relatively undeformed granitic and hybrid rocks of the Sybella Batholith. The question of whether myrmekite grows in the solid-state or within a melt (or a residual fluid; Hibbard 1979) and the effect of strain on the growth or nucleation of myrmekite (Vernon 1991b) are still debatable. However, it is clear that myrmekite is related to grain boundaries or to fractures within the grain, and
with the implication that the feldspar matrix was largely established before its growth. Myrmekite occur as lobes that project into the K-feldspar phenocrysts (typically microcline) from the margin, and along fractures and grain boundaries of plagioclase inclusions within the phenocryst. High-temperature orthoclase contains dissolved Na and Ca and at low temperatures the orthoclase, under stress, inverts to microcline and exsolves the Na and Ca to the margin of the crystal to form myrmekite. Because calcic plagioclase requires less silica in its lattice than in K-feldspar or sodic plagioclase, silica is left over to form quartz vermicules in the myrmekite that is proportional to the Ca-content. In phenocrysts, myrmekite is preferentially developed in areas with localized stress and strain, occurring in zones parallel to the foliation development (Vernon 1991b; Simpson & Wintsch 1989). However, myrmekite may completely surround the phenocryst and also occur on fractures in areas which apparently did not have high compressive stress (or strain) across them. Progressive deformation destroys the delicate structure of myrmekite, and leads to the development of fine-grained folia. The growth of myrmekite is not controlled by the strain energy, but rather provided access for fluid movement.

3.4 ESTIMATES OF INTENSIVE VARIABLES

Various experimentally calibrated methods have been developed, utilizing coexisting mineral assemblages, to determine the intensive parameters of crystallization (P, T, \( f_O^2 \) and \( f_{H_2O} \)) of granitic magmas. Of these methods, very few are applicable to rapakivi A-type magmas, the main difficulty being that they usually consist of disequilibrium mineral assemblages (i.e. hornblende and biotite crystallizing later than K-feldspar and quartz). Moreover, the minerals commonly show re-equilibration during the slow cooling under plutonic conditions (Ramo & Haapala 1995). Analyses of amphiboles used in estimating P-T conditions of emplacement are listed in Table 3.2.

3.4.1 Pressure
The composition of hornblende is thought to be depth dependent. Pressure estimates were calculated for the granites using the Al-in-hornblende barometer of Johnson & Rutherford (1989) in addition to that of Anderson & Smith (1995). Calculations of
pressures using this technique for intrusions with the appropriate equilibrium assemblage (plagioclase + hornblende + quartz + biotite + orthoclase/sanidine + titanite + Fe-Ti oxides + melt + fluid phase; Hammerstrom & Zen 1986) are listed in Table 3.3. Anderson & Smith (1995) caution against the use of existing Al-in-hornblende barometers for low $f_O^2$, and granites with amphiboles having Fe/(Fe+Mg) above 0.65 as they are outside the range of experimental calibrations. Increased temperature and low $f_O^2$/high Fe content show a positive effect on total Al content in hornblende, hence producing anomalously high estimates of pressure determinations. Anorogenic granites commonly have high-Fe and high-temperatures and the application of the Al-in-hornblende barometers needed to be used judiciously. As the melts are also undersaturated in H$_2$O, a fluid phase may not be present, which is required for the equilibrium mineral assemblage of Hammerstrom & Zen (1986).

Using the geobarometer of Johnson & Rutherford (1989) hornblende crystallized at 3.8 kbar (hybrids) to 5.9 kbar (Main Phase granite). The Main Phase granite pluton is high temperature and has very Fe-rich mafic minerals, including hornblende with Fe/(Fe+Mg) ratios of 0.85 (Figure 3.3), which is similar to that of the whole rock. Main Phase hybrids, rapakivi granitoids have lower Fe/(Fe+Mg) ratios of <0.65 and higher $f_O^2$ that fall within the experimental calibrated range for geobarometer based on hornblende crystallization. Therefore, a pressure of approximately 4 kbar may be a reasonable estimate.

In recognition of the effect of temperature, pressures ranging from 1.5 kbar for hybrids to 3 kbar for the Main Phase granite were determined using the calibration of Anderson & Smith (1995). Discrepancies could also result in part from partial early crystallization of the Main Phase magmas under high-pressure conditions, followed by emplacement as a crystal rich magma at lower pressures. The lower pressure determined for the rapakivi granitoids could reflect the interaction at emplacement level of the granite and mafic diorites.

Re-equilibration of hornblende during later Isan Orogeny high T-low P metamorphism may also have effected the composition and pressures determined. Independent estimates of pressure may be able to be derived from the metamorphic assemblages.
### Table 3.2  Amphibole analyses used estimating P-T conditions of emplacement. An content from coexisting plagioclase; Fe$_2$O$_3$ and FeO ratios calculated using the method of Holland & Blundy (1994); ed-tr and ed-ri geothermometers of Holland & Blundy (1994) at 3 kbar pressure; biotite temperatures calculated using Wones (1981).
### Table 3.3  Al-in-hornblende Geobarometry for the different rock types.

<table>
<thead>
<tr>
<th>Sample Id</th>
<th>Estim. Temp</th>
<th>Al Total</th>
<th>equation 1</th>
<th>equation 2</th>
<th>equation 3</th>
<th>equation 4</th>
<th>equation 5</th>
</tr>
</thead>
<tbody>
<tr>
<td>Error (kbars)</td>
<td>3.00</td>
<td>1.00</td>
<td>0.22</td>
<td>0.60</td>
<td>0.60</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

#### Equations

1. \( P (\pm 3 \text{ kbar}) = -3.92 + 5.03(A\text{\textsubscript{T}}); r^2 = 0.80 \) (Hammerstrom & Zen, 1986)
2. \( P (\pm 1 \text{ kbar}) = -4.76 + 5.64(A\text{\textsubscript{T}}); r^2 = 0.97 \) (Hollister, 1987)
3. \( P (\pm 0.22 \text{ kbar}) = -3.46 (\pm 0.24) + 4.23 (\pm 0.13) (A\text{\textsubscript{T}}); r^2 = 0.99 \) (Johnson & Rutherford, 1989)
4. \( P (\pm 0.6 \text{ kbar}) = -3.01 + 4.76 (A\text{\textsubscript{T}}); r^2 = 0.99 \) (Schmidt, 1992)
5. \( P (\pm 0.6 \text{ kbar}) = 4.76*\text{Al}^T - 3.01 -([T (^\circ \text{C}) - 675]/85) * \{0.530* \text{Al}^T + 0.005294 *[T (^\circ \text{C}) - 675]\} \) (Anderson & Smith 1995)

Assuming a constant crustal density of ~2.75, each kbar of pressure equates to approximately 3.4 km depth.
within the adjacent country rocks of the May Downs Gneiss. Prior to this investigation, no thermal aureole had been recognized. Unraveling the complicated metamorphic history cordierite-biotite andalusite/K-Feldspar-sillimanite gneiss and correlating it with the current understanding of deformation history of the area has been based on several assumptions and also new evidence:

(1) that the early deformation and metamorphism (early cordierite growth that grew within a steep foliation, previously termed early S2; Rubenach 1992) within the May Downs Gneiss occurred pre-syn emplacement of the Sybella Batholith, and does not represent basement of Baramundi Orogeny age. Evidence for this includes rafts of folded gneiss found within the granite near margins and undeformed phases of granite cross-cutting strongly deformed May Downs Gneiss.

(2) that the later, second deformation and metamorphism (andalusite and sillimanite growth, late S2; Rubenach 1992) recorded in the May Downs Gneiss was related to the Isan orogeny.

Using the assumption that the metamorphic conditions as determined within the May Downs Gneiss reflect the emplacement of the Main Phase granites, pressure would then be approximately 4 kbars (Rubenach 1992; Rubenach pers. com. 2002).

3.4.2 Temperature

Estimates of magmatic temperatures can be obtained from consideration of the solubility of apatite and zircon in the melt to approximate the temperature of their saturation (Watson & Harrison 1983; Harrison & Watson 1984). These estimates are based on models of the temperature of apatite and zircon saturation using P2O5 and Zr concentrations in the granitoids. They assume that the concentrations of P2O5 and Zr in the magma reflects the original composition of the magma, that no apatite or zircon was added to the melt through cumulate, restite or country rock assimilation, and that the thermometers are only valid if the rock represents a liquid composition. There is no zircon Pb isotopic or petrographic evidence for inherited zircons in the Main Phase granite (Connors & Page 1995) or microgranite (Coleborn 1999). Mineral saturation temperatures for the different phases within the Sybella Batholith were interpreted in view of the proposed mixing history, and it is unclear whether these temperatures represent true mineral saturation or whether they reflect inheritance from end members.
As zircon occurs as euhedral inclusions in biotite and groundmass minerals in the Main Phase granites, they are interpreted to have formed after K-feldspar crystallization and not early in the crystallization of the felsic intrusions. Therefore, saturation temperature estimates will approximate the temperature during crystallization and not initial melt temperature (Figure 3.24a: ~880°C for Main Phase granite and 860-800°C for the microgranites). Zircons have an anhedral to skeletal morphology in the rapakivi granitoid hybrids and hence the thermometer is inapplicable to these rocks. Apatite saturation in mafic magmas is commonly used to estimate temperatures, however as the apatite was found to be acicular within the groundmass (due to rapid crystallization resulting from chilling), rather than euhedral inclusions in early-formed phases, it is thought that the mafic magmas were under-saturated with respect to P$_2$O$_5$. This thermometer was found to be invalid and generally estimated low temperatures for the mafic units (Figure 3.24b). Some of the mafic units may have achieved apatite saturation with temperatures estimated at approximately 1020°C.

Estimates of solidus conditions using the hornblende-plagioclase geothermometry technique of Holland & Blundy (1994) are listed in Table 3.2 and temperatures are typically in the range of 855-900°C for the Main Phase granite, 830-800°C for the rapakivi granitoid hybrids. Lower temperatures (720-770°C) were estimated for the microgranites and associated hybrids. The temperature estimated for the mafic diorite (Mafic Hybrid Complex) using hornblende-plagioclase geothermometry was low (720°C), however the amphibole analyzed was petrographically determined to be formed by the alteration of pyroxene in a subsolidus reaction. Ferric iron has been incorrectly calculated in some samples, and the use of the silica-saturated thermometer has led to anomalously high calculated temperatures for the Main Phase granites, hence the Si undersaturated thermometer (Equation B of Holland & Blundy 1994) was preferred.

Another measure of the solidus temperature of the granitoids is the titanium content of biotite, which is recognized to increase with temperature when buffered by the presence of magnetite and a Ti-bearing phase (ilmenite or titanite; Wones 1981). Based on the titanium and iron content of biotite, TiO$_2$ decreases (5.5 to 1.2 wt%) with increasing SiO$_2$ of the host rock. Temperature estimates using this procedure suggest that biotite
Figure 3.24  (a) Zircon saturation temperature for granitoids (Harrison & Watson, 1984); (b) Apatite saturation (Watson & Harrison, 1983); (c) comparison of hornblende-plagioclase geothermometry for quartz saturated vs unsaturated temperature (Holland & Blundy, 1994); (d) Temperature with calculated Fe$^{3+}$/Fe$^{2+}$ ratio (Holland & Blundy, 1994).
last equilibrated at 350-450°C in the Main Phase granite, at 505-520°C in the rapakivi granitoid hybrid, at 630-670°C in the mafic hybrid, and at 680-750°C in the diorite within the Mafic Hybrid Complex (Table 3.2). Temperature estimates from biotite within the microgranite hybrids group at 450-485°C and 640-697°C, and at 630-750°C in the Mosses Tank Dolerite and mafic hybrids. This suggests that biotite may have reequilibrated at subsolidus temperatures in all rock types.

3.4.3 Oxygen Fugacity

The oxygen fugacity of a magma can be either intrinsic to the source region or developed during the history of the magma. It is difficult to establish original oxygen fugacities of primary magmas because oxidation can occur during the slow cooling and crystallization of magma (e.g. Czamanske & Wones 1973). The assemblage titanite + magnetite + quartz in granitic rocks permits an estimate of relative oxygen fugacity (Wones 1989). The presence of euhedral titanite and magnetite as early crystallizing phases in silicic rocks would indicate that the initial magma was relatively oxidized, however if the titanite was formed during late crystallization it may have resulted from late-stage oxidation. The occurrence of magnetite is insufficient information to determine the oxygen fugacity because it can form under low $fO_2$ conditions.

Biotite and hornblende Fe/(Fe+Mg) ratios are controlled by intensive parameters (Wones 1981), particularly the $fO_2$ which exerts the strongest control on mafic silicate chemistry. With increasing $fO_2$, the Fe/(Fe+Mg) ratio of silicates markedly decreases independent of the whole rock chemistry (Wones 1989). Amphiboles occurring in the Main Phase granites range from Fe-rich (hastingsite) to more Mg-rich (ferro-edenite) in the rapakivi granitoids and the Mafic Hybrid Complex, to ferro-hornblende in the microgranite hybrids. The rocks of the Mafic Hybrid Complex and rapakivi granitoids have significantly lower Fe/(Fe+Mg) ratios (Figure 3.25a), which is interpreted to be oxidation associated with interaction of mafic with felsic magmas (and regionally coincident with a strong magnetic anomaly). The calculated Fe$^{3+}$/($Fe^{3+}$+Fe$^{2+}$) ratios for the amphiboles are low and fall in the range of 0.10 to 0.28, with the Main Phase granite at approximately 0.23. This indicates that the magmas initially crystallized under low
$fO_2$ conditions (Anderson & Smith 1995) that was near to or below the QFM (quartz-fayalite-magnetite buffer).

The biotite-titanite±hornblende mineralogy of the Main Phase granite, the late crystallization of the hydrous phases and the Fe-Ti oxide mineralogy, constitutes a magnetite series, and implies low water content and an elevated oxygen fugacity. The magnetite in the Main Phase granite is commonly euhedral and occurs within the K-feldspar phenocrysts and groundmass. The presence of anhedral titanite, magnetite and quartz in the Main Phase granite indicates that the magma was oxidized (Wones 1989) relatively late in the crystallization history. Estimates using the equilibrium expression (see below) of Wones (1989) suggest the magma had an $fO_2$ of $10^{-11}$ to $10^{-12}$ (between 900 to 850°C respectively, and at 4 kbars) (Figures 3.25bc). These values are thought to represent late stage magmatic conditions rather than the initial oxygen fugacity.

Titanite + magnetite + quartz = hedenbergite + ilmenite + $O_2$

$$\log fO_2 = -(30939/T) + 14.98 + [0.142 \times (P-1)/T]$$

Where $T = \text{temperature (K)}$; $P = \text{pressure (bars)}$. 

Although the Main Phase granite is magnetite series, the very high Fe/(Fe+Mg) ratios (~0.85) reflect crystallization at relatively low $fO_2$, below QFM (Anderson & Smith 1995). The initial magma was thought to be at or below QFM, and experienced a later stage of oxidation, which is reflected in the titanite, magnetite and quartz assemblage 1-2 log units above the QFM.

Constraints on the $fO_2$ of the hornblende-absent microgranites that contain > 70 wt% SiO$_2$, titanite, magnetite, and quartz are not as well defined, but given that magnetite is present, the $fO_2$ of the melt must have remained below the hematite-magnetite buffer.

In the mafic and intermediate units of the Mafic Hybrid Complex, the ilmenite lamellae in the pyroxenes and ilmenite within the early biotite suggest that the mafic magma was initially very low in $fO_2$. Mixing of this reduced mafic magma and a more oxidized magma would lead to the formation of magnetite, and this process would explain the
Figure 3.25  Oxygen fugacity.
(A) Estimated oxygen fugacity based on $\text{Fe}^{2+}/(\text{Fe}^{2+}+\text{Mg})$ content of amphiboles. Dashed lines are from Anderson & Smith (1995). (B) Calculated oxygen fugacity based on the assemblage magnetite+titanite+quartz (Wones, 1989) with mineral assemblage of the main phase represented by reactions 2, 3, 7, and 7a; (C) reaction of hybrids during oxidation (Wones, 1989).
occurrence of both magnetite and ilmenite in the hybridized dolerites, diorites and rapakivi granitoid hybrids. The presence of abundant late stage anhedral titanite in the hybrids is possibly the result of oxidation related to mixing the Main Phase granite and dolerite (see reaction below).

\[
\text{Pyroxene + ilmenite = amphibole + magnetite + quartz + titanite}
\]

The highest $f_O^2$ occurs in the Mafic Hybrid Complex, and is characterized by extensive aeromagnetic anomalies (Figure 2.4-5). The hornblende in these rocks has Fe/(Fe+Mg) ratios in the range of 0.6 to 0.4 (Figure 3.25a). Mafic units that are hybridized with granite show the above reaction, which indicates a change from a reduced ilmenite-bearing magma to a relatively oxidized magnetite-bearing magma.

### 3.4.4 Water Fugacity

Water content of a melt strongly influences the magma’s chemical and physical properties, including viscosity and crystallization. The viscosity of a melt decreases by several orders of magnitude when the water content increases from 0 to 2 wt. %, whereas a small decrease is observed from 6 to 8 wt. % (Holtz et al. 1996).

The early crystallization of biotite (directly after augite and before hornblende) in the mafic units is indicative of water-undersaturated magmas (Wones & Gilbert 1982). The occurrence of biotite and hornblende as anhedral grains that fill intertices between quartz and feldspar crystals, indicates that the main intrusive granite phases also crystallized from H$_2$O-undersaturated magmas. Other features of the granites that include the paucity of pegmatites, late crystallization of hydrous phases, relatively shallow emplacement and high temperatures of crystallization, suggest that conditions remained largely anhydrous throughout crystallization of this suite.

For high-temperature melting reactions (850-900° C) at 5-7 kbars (Qtz-Or-Ab normative source depth estimation; Figure 3.26a) that involve biotite and/or breakdown reactions, the minimum water content is estimated to range from 2.4-4.4 wt%, although it may be close to 4.5 wt% (Figure 3.26b; Holtz et al. 2001), which is similar to other high temperature A-type granites (Dall'Agnol et al. 1999). Melting to form the microgranites at 7-9 kbar at lower temperature would suggest the melt had higher water content (Figure 3.26b).
Figure 3.26
(A) Qtz-Or-Ab normative mineralogy for source depth estimation.
(B) water content at estimated source depth of 5-7 Kbars and 800-900 deg C (Holtz et al 2001).
3.5 SUMMARY

The association between the Main Phase granites and Mafic Hybrid Complex, and between the microgranite and Mosses Tank Dolerite, has produced rock units that have textures indicative of mixing. Mineral disequilibrium textures, compositional variation, textures and mineralogy reflect the change in temperature, water and oxygen fugacity. Macroscopic evidence for widespread mixing within the Mafic Intrusive Suite is supported at a microscopic scale, as is the interaction between the Mafic Intrusive Suite and both the Main Phase granite and microgranite. Besides the conspicuous rapakivi texture in the Main Phase granite, the granites exhibit other petrographic characteristics generally associated with A-type granite suites. The Main Phase granite typically shows little compositional variation and is characterized by Fe-enriched mafic silicates, high fluorine, and alkali feldspar content that is similar to other rapakivi granites throughout the world (Ramo & Haapala 1995).

Hybridization of granitic and dioritic magmas and the exchange of partly crystallized mineral phases (especially feldspars and quartz) can account for the mineralogical and textural features of the intrusions that are interpreted as hybrids, including the rapakivi granitoids, Mafic Hybrid Complex, and rocks of an intermediate composition at mafic-felsic rock intrusive contacts. During mixing, the two magmas exchanged early-crystallized plagioclase crystals (Section 3.3.2). In the intermediate hybrids, early-formed plagioclase crystals (high An) captured from a more mafic magma (such as the MHC or MTD) would have been reheated and partly resorbed before being overgrown by andesine. More mafic hybrids from the MHC have reverse zoned plagioclase with a low An core and a higher An rim. Compositional spikes and inclusion-rich zones in feldspars are also indicative of mixing and associated compositional and temperature changes in the magma. Mantle quartz xenocrysts in the diorite and hybrid magmas indicate crystal exchange and mineral disequilibrium during interaction and mixing of partly crystallized melts. The presence of acicular apatite in the hybrids points to a high nucleation rate triggered by undercooling of the dioritic and doleritic magmas during mixing (Castro et al. 1991 and Hibbard 1991).

Feldspar megacrysts in the Main Phase granites are not in textural equilibrium with the host rock, as evidenced by plagioclase rims on K-feldspar megacrysts, and K-feldspar
with euhedral cores and inclusion rich rims. This indicates that much of the Main Phase granite is characterized by disequilibrium textures that possibly resulted from a combination of mafic and felsic magma interaction, high fluorine content and changes in P-T-fO₂-fH₂O conditions.

Of the variously calibrated methods for determining the intensive parameters of granitic magma crystallization (T, P, fO₂, fH₂O), very few are applicable to rapakivi A-type magmas. The main difficulty is that the granites consist of disequilibrium mineral assemblages (different generations and order of mafic/silicic minerals etc) and the high Fe-content of minerals. However, temperatures of 850-900°C and approximate emplacement pressure of 4 kbars for the Main Phase granite are similar to other A-type granites world-wide (e.g. Anderson & Bender 1989; Smith et al. 1997). The absence of source rock restites is also indicative of high magma temperatures. The pressure estimates determined on the basis of Al-in-hornblende show discrepancies with a difference of 1.5 kbar between the Main Phase granites and the rapakivi granitoid hybrids, which possibly reflects inherent problems with the geobarometers used. The groundmass of quartz K-feldspar porphyritic dykes represents rapid in situ crystallization at the level of emplacement, and possibly lower emplacement pressures than the main intrusive suites.

The granites are directly indicative of the oxidized state of the source, therefore are interpreted as having relatively low initial oxygen fugacities. The high Fe⁹/Mg ratios of mafic minerals and low Fe³⁺ (calc.) in minerals indicate low fO₂ in rocks. Low fO₂ followed by an oxidation event (possibly associated with a mixing event) is recorded by the higher fO₂ levels determined for the Mafic Hybrid Complex and rapakivi granitoids (characterized by abundant magnetite and lower Fe content of the mafic minerals within the hybrids) and Main Phase suite of rocks (with late anhedral titanite, magnetite and quartz). The occurrence of large biotite flakes ± late amphibole in the mafic units, and the apparent late crystallization of mafic minerals in the granites indicates that the magmas were initially undersaturated in H₂O. The low water fugacities and high temperature of the melts enabled the magmas to intrude into the upper crust at depths of 8-17 km or shallower.
CHAPTER 4
GEOCHEMISTRY

4.1 INTRODUCTION

Geochemical variation in suites have been ascribed to several processes such as magma mixing or mingling assimilation, fractional crystallization, varying degrees of partial melting and separation of unmelted source material or restite from a melt, composition of source, and various combinations of those processes. The aim of this section is to geochemically affirm the suggested mixed and mingled interaction between the mafic and felsic intrusions within the northeastern portion of the Sybella Batholith, and to propose an appropriate petrogenetic model for their overall development. Variation at outcrop scale within the mafic suite of rocks is also evaluated regarding geochemical processes. In order to test the mixing hypothesis, the geochemical analyses of samples thought to be directly related by mixing (determined from field relationships; Chapter 2) are closely evaluated.

The different rocks analyzed were grouped into seven categories of igneous intrusions with three groups associated with microgranite in the northern region, Kitty Plains, four groups associated with Main Phase granite in the Guns Knob and Easter Egg regions, and two categories of country rocks, based primarily on field relationships and petrographic characteristics. The groups include:

- (1) Mosses Tank Dolerite; (2) Microgranite Hybrid; (3) Microgranite;
- (4) Mafic Hybrid Complex; (5) Rapakivi Granitoid Hybrids; (6) Main Phase Granite; (7) Porphyritic granite and aplite;
- Country Rocks - Eastern Creek Volcanics: (1) amphibolite and (2) mafic gneiss.

Thirty-seven, fresh, 10-20 kg samples were prepared for geochemical analysis through this study, with an additional seven samples from Kitty Plain (Coleborn 1999), thirteen from Easter Egg region (Sisois 2000), and ten mafic gneiss and amphibolite samples from the May Downs Gneiss region (Gunter 1996) where there has been geological (field relationships) and petrological control on rock type. Sample descriptions and
locations, geochemical analysis from this study and selected additional analysis from the region, are given in Appendix IV, V and VI respectively. Each sample was crushed using a hydraulic press and ground in two aliquots using Cr-steel and W-carbide mills. Major components and some trace elements were determined by X-ray fluorescence spectrometry on glass discs (major elements) and pressed powder pellets at the Advanced Analytical Centre, James Cook University. LOI was calculated by measuring the relative mass loss (e.g. mineral H₂O and CO₂) of a ‘dry’ (pre-heated to ~100°C for 5-8 hours) powdered sample after heating it to 1000°C for <12 hours. Rare-earth elements, Co, Ta, Hf, Th and U were determined by neutron activation analysis by Becquerel Laboratories Pty Ltd, Lucas Heights, NSW, Australia.

4.2 COMPOSITIONAL DISTINCTIONS AMONG INTRUSIVE UNITS

Major and trace element compositions for dolerites, granites and hybrids are plotted on selected graphs in Figure 4.1 - 4.4. Most analyses form coherent arrays on major element Harker diagrams (Figure 4.1), although some show considerable scatter. The mafic intrusives (Mafic Hybrid Complex and Mosses Tank Dolerites) are metaluminous, subalkaline rocks that range from 45-55 wt% SiO₂, and have medium-high K (Le Maitre 1989) and within-plate (Pearce & Cann 1973) tholeiitic affinities (Irvine & Barager 1971; Figure 4.4). The more mafic rocks of the rapakivi granitoid hybrids (quartz diorites) are metaluminous, subalkaline rocks contain high-K (with one sample medium-K), and range from 58.2 - 65 wt % SiO₂. The texturally distinct phenocrystic/xenocrystic intrusions of the rapakivi granitoid hybrids range from 63-69 wt % SiO₂, are high-K, metaluminous, subalkaline granodiorite to monzodiorite. The SiO₂ composition of this group overlaps with the Mafic Hybrid Complex. They have major and trace element compositions intermediate between the mafic and granite phases. The Main Phase felsic granitoids are metaluminous, subalkaline monzogranites, are high-K, iron-enriched and have approximately 70 wt % SiO₂ (Figure 4.4). These rocks typically have ‘within-plate’ (Pearce et al. 1984) and ‘A-type’ (Whalen et al. 1987) granite affinities on discrimination diagrams. The alkaline enriched but not peralkaline geochemical character of the Main Phase granites is typical of many rapakivi related granite associations (Ramo & Haapala 1995). The microgranites are
### Table 4.1 Geochemical analysis of intrusive units from the Sybella Batholith

<table>
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<tr>
<th>Sample</th>
<th>MAFIC HYBRID COMPLEX &amp; RAPAKIVI HYBRIDS</th>
<th>MAIN PHASE GRANITE SUITE</th>
<th>negative numbers below detection level</th>
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<td>EE Dolerite</td>
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<td>0.16</td>
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| Cl | 620 | 270 | 420 | 400 | 590 | 510 | 540 | 670 | 500 | 420 |
| Ba | 650 | 584 | 675 | 1209 | 1245 | 1213 | 928 | 1094 | 860 | 1257 |
| Rb | 107 | 87 | 109 | 90 | 159 | 103 | 168 | 262 | 261 | 178 |
| Sr | 280 | 237 | 243 | 332 | 319 | 228 | 218 | 168 | 185 | 213 |
| Pb | 9 | 11 | 9 | 15 | 14 | 18 | 16 | 22 | 23 |
| Th (naa) | 6.02 | 10.10 | 12.00 | 8.49 | 19.90 | 12.10 | 18.80 | 36.70 | 26.60 | 21.30 |
| U (naa) | -2.00 | -2.00 | -2.00 | -2.00 | -2.00 | -2.00 | -2.00 | -2.00 | -2.00 | -2.00 |
| Zr | 291 | 277 | 218 | 434 | 401 | 401 | 331 | 572 | 299 | 293 |
| Nb | 15 | 18 | 16 | 22 | 28 | 34 | 27 | 38 | 22 | 23 |
| La (naa) | 53.90 | 55.60 | 51.50 | 89.00 | 94.70 | 86.50 | 100.00 | 159.00 | 78.70 | 93.50 |
| Ce (naa) | 105.00 | 115.00 | 103.00 | 178.00 | 182.00 | 190.00 | 193.00 | 269.00 | 152.00 | 183.00 |
| Sm (naa) | 9.06 | 9.48 | 8.88 | 13.50 | 15.10 | 14.90 | 13.60 | 14.50 | 11.40 | 12.80 |
| Eu (naa) | 1.89 | 2.12 | 1.76 | 2.59 | 2.92 | 3.20 | 2.41 | 2.08 | 1.95 | 2.15 |
| Th (naa) | 1.25 | 1.33 | 1.30 | 1.95 | 1.91 | 2.07 | 1.86 | 1.67 | 1.66 | 1.90 |
| Yb (naa) | 3.36 | 4.57 | 3.58 | 5.26 | 5.40 | 6.25 | 6.08 | 5.02 | 4.82 | 5.20 |
| Lu (naa) | 0.44 | 0.60 | 0.48 | 0.70 | 0.75 | 0.82 | 0.81 | 0.64 | 0.65 | 0.64 |
| Sc (naa) | 29.30 | 36.40 | 30.40 | 27.10 | 22.50 | 24.90 | 20.80 | 16.00 | 18.80 | 13.20 |
| V (naa) | 217 | 275 | 220 | 194 | 149 | 92 | 113 | 88 | 117 | 60 |
| Cr (naa) | 196.0 | 14.8 | 91.3 | 93.6 | -5.0 | -5.0 | -5.0 | 10.2 | 58.9 | 11.8 |
| Co (naa) | 42.70 | -100.00 | 39.80 | 24.00 | -10.00 | 11.00 | 30.00 | 32.00 | 32.00 | 32.00 |
| Ni (naa) | 77.36 | 60 | 51 | 6 | 14 | 21 | 11 | 18 | 11 | 74 |
| Cu | 68 | 67 | 58 | 72 | 34 | 74 | 38 | 42 | 30 | 32 |
| Zn | 93 | 112 | 84 | 112 | 95 | 118 | 80 | 116 | 53 | 39 |
| Ga | 19 | 23 | 21 | 22 | 21 | 21 | 21 | 24 | 19 | 20 |
| Hf (naa) | 7.05 | 7.39 | 6.74 | 10.80 | 11.40 | 11.60 | 7.22 | 18.10 | 8.80 | 8.67 |
| Fe* | 0.62 | 0.72 | 0.66 | 0.73 | 0.84 | 0.84 | 0.81 | 0.81 | 0.72 | 0.81 |
| As* | 0.83 | 0.84 | 0.81 | 0.85 | 0.83 | 0.83 | 0.84 | 0.84 | 0.85 | 0.83 |
| Cr | 0.36 | 0.37 | 0.27 | 0.27 | 0.27 | 0.35 | 0.96 | 1.08 | 0.95 | 0.95 |
| Ba | 0.16 | 0.15 | 0.16 | 0.07 | 0.13 | 0.08 | 0.20 | 0.24 | 0.23 | 0.14 |
| K/Rb | 167 | 177 | 165 | 236 | 180 | 203 | 165 | 139 | 178 | 217 |
### MOSSES TANK DOLERITE

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Table 4.1 Cont. Geochemical analysis of intrusive units from the Sybella Batholith

**Notes:** negative numbers below detection level
Figure 4.1  Variation of major element contents of rocks from the Sybella Batholith. (a) SiO₂-TiO₂; (b) SiO₂-CaO; (c) SiO₂-Al₂O₃; (d) SiO₂-Na₂O; (e) SiO₂-Fe₂O₃T; (f) SiO₂-K₂O; (g) SiO₂-MgO; (h) SiO₂-P₂O₅.
Figure 4.2  Variation of selected trace element contents of rocks from the Sybella Batholith. (a) SiO$_2$-Sr; (b) SiO$_2$-Rb; (c) SiO$_2$-Ba; (d) SiO$_2$-Zr; (e) SiO$_2$-V; (f) SiO$_2$-Y; (g) SiO$_2$-Pb; (h) SiO$_2$-Nb.
Figure 4.3  Selected trace and major element variation diagrams: (a) SiO$_2$-Cr; (b) SiO$_2$-Ni; (c) SiO$_2$-F; (d) SiO$_2$-Cl; (e) MgO-K$_2$O; (f) MgO-CaO; (g) Rb-Ba; (h) Rb-K/Rb.
Chapter 4

Figure 4.4 Tectonic geochemical affinities: (a) subalkaline trend (Irvine & Barager 1971); (b) metaluminous mafic and granite suite with microgranites transitional to peraluminous (Maniar & Piccoli 1989); (c) tholeiitic suite (Irvine & Barager 1971); (d) chemical classification; (e) granite discrimination diagrams of Whalen et al (1987) and (f) Pearce et al (1984) illustrate the main phase granites as within-plate granites, whereas the microgranites and the porphyritic granites trend into the field for volcanic arc granites, and fractionated granite suites; (g) within-plate basalt (Pearce & Norry 1979); and (h) shows the distinction between the Mosses Tank Dolerites which plot in the “within plate basalt” field, and the Mafic Hybrid Complex which have increasing Zr contents possibly due to granite mixing (Pearce & Cann 1973). Symbols as in Figure 4.1.
subalkalic, and span the metaluminous and peraluminous boundary (Figure 4.4b). They are more closely related to ‘fractionated-felsic’ granites and have the highest SiO₂ with 75-77 wt %, although comparable with the highly fractionated porphyritic granite of Main Phase association. This could reflect either a different source for the microgranites or the effect of Y depletion during differentiation.

The mafic, felsic and hybrid granitoids have very similar profiles on a primitive mantle-normalized REE plot (Figures 4.5 - 4.9). All granitoids display a negative Eu anomaly and are Y undepleted and enriched in Th and LREE. The rocks have a within-plate compositional affinity and the magmas were probably derived from a mixed mantle-lower crustal source that was emplaced in an extensional setting.

4.2.1 Mafic Rocks

The mafic rocks of the Sybella Batholith are divided into two groups based primarily on geochemical properties, which is supported by field relationships. The first group consists of the Mosses Tank Dolerites (MTD) with low SiO₂ (47.8-53.9 wt %). The second group of mafic rocks is the Mafic Hybrid Complex (MHC), which has higher SiO₂ content (51.1-55.1 wt%) and includes the highly magnetic diorites associated with the Main Phase granite. Some major oxide components (TiO₂, MgO, Al₂O₃, Na₂O, and P₂O₅) have scattered trend with increasing SiO₂ (Figure 4.1). In the mafic rocks CaO, Fe₂O₃T, and MgO decrease, whereas K₂O and P₂O₅ increase with increasing SiO₂. Trends within this scattered group may show TiO₂ and Na₂O increase, and Al₂O₃ decrease with SiO₂. The Mg# ranges from 0.37 to 0.27 for both of the groups. The Mosses Tank Dolerites have a higher CaO content, and lower K₂O content (Figure 4.3ef) than the Mafic Hybrid Complex. On the AFM, Ti-Zr-Y and Zr/Y-Zr diagrams (Figure 4.4), the distinction between the two groups is also apparent. Aside from the distinction between the Mosses Tank Dolerite and Mafic Hybrid Complex, there is high/low Ni, Cr and MgO clustering of compositions, irrespective of what group they belong to (Figure 4.3ab).

On multi-element primitive mantle (Sun & McDonough 1989) normalized diagrams (Figure 4.5) both groups of mafic rocks show a strong depletion in Nb and Sr. The REE contents of the MTD are lower than the higher SiO₂ MHC, and has little to no Eu
Figure 4.5  Primitive mantle normalized multi-element and REE pattern of Mosses Tank Dolerites, diorites and enclaves of the Sybella Batholith. Symbols as in Figure 4.1.
anomaly with a flat La-Lu trend (Figure 4.5). MHC rocks have a “stacked” REE trend with increasing contents of most elements with increasing SiO\textsubscript{2}. The mafic sample with the lowest SiO\textsubscript{2} content (IS-17: 51.1 wt \%) has a REE and multi-element pattern similar to the MTD, however it has a pronounced Eu anomaly where the MTD does not. Mafic enclaves found as pillows within the Main Phase granite have similar REE contents to the units within Mafic Hybrid Complex. The mafic enclave from Kitty Plain has lower REE content than the other dolerites from that region (MTD) and a higher SiO\textsubscript{2} content.

4.2.2 Main Phase Potassic Rocks

The Main Phase granites have a narrow silica range with SiO\textsubscript{2} contents of 69.6-70.5 wt\% (Figure 4.1). This narrow range of composition is reflected on REE and multi-element variation diagrams (Figure 4.6). A range of aplite and porphyritic granites associated with the Main Phase granites are 74-75 wt\% SiO\textsubscript{2} and lie within the A-type granite fields (Figure 4.4) similar to the Main Phase granite. The LREE are enriched compared to the Main Phase granite and depleted in HREE. They also have an increased Eu anomaly. The Main Phase granite sample (520) from the northern margin of Queen Elizabeth Pluton has a higher SiO\textsubscript{2}, REE content, and trace elements (Nb, Y and Ce) and lower V and Ti than the other granites and lies distinctively away from the group on binary plots and discrimination diagrams (Figure 4.4 e,f).

In general, the potassic granitoids show a depletion of Ba, Nb, Sr, Eu and Ti, and enrichment in Th, U and LREE on the primitive mantle (Sun & McDonough 1989) normalized element diagrams (Figure 4.6). These trends typically increase in magnitude with SiO\textsubscript{2} content from Main Phase to the aplite/porphyritic granite samples. Rb/Sr ratios increase from mafic to felsic, however they are notably higher in the microgranites than that of either the Main Phase or the fractionated porphyritic granite (Table 5.1). Rb/Ba increases with increasing SiO\textsubscript{2} from Main Phase to the aplite/porphyritic granite – with the exception of sample 520 from the Queen Elizabeth Pluton – a feature consistent with crystallization of feldspar.
4.2.3 Microgranites

Samples of the microgranites are richer in SiO$_2$ than the Main Phase granites (Table 5.1) and contain lower concentrations of rare earth elements (Figure 4.6). Due to the low LREE abundances of the microgranites, they are more closely related to ‘fractionated-felsic’ granites (Figure 4.4e). The high SiO$_2$ content (75-77 wt%) is comparable with the highly fractionated porphyritic granite of Main Phase association; the distinction is highlighted by the differences in the concentration of Zr, Ba, Nb, Y, Pb and Rb and Rb/Sr ratios (Figure 4.6). Low TiO$_2$, Fe$_2$O$_3$, MnO, MgO and CaO contents also characterize the phase. The microgranites form a chemically distinct group compared with those of the Main Phase and fractionated porphyritic granite.

The microgranite that brecciated the mixed unit at Mica Creek is distinctively different from the other granite phases within the Sybella batholith, with the REE and multi-element diagrams showing crossed-trend lines (Figure 4.6).

4.2.4 Granite-Dolerite Hybrids

Field evidence and petrographic features indicating chemical and/or thermal disequilibrium between the groundmass minerals and phenocrysts (e.g. corroded and mantled phenocrysts of feldspars and quartz) suggests interaction, and to some degree, mixing between silicic and basic magmas. There are two distinctive groups of rocks interpreted to be hybrids: those that resulted from mixing Mosses Tank Dolerite with the microgranite, and those that mixed Mafic Hybrid Complex (dolerite to diorite) with the Main Phase granite. The hybrids have SiO$_2$ contents from 58.2 wt% (intermediate hybrids) to 68.7 wt% (granite hybrids), which is intermediate between the mafic and felsic end-members.

RAPAKIVI GRANITOID HYBRIDS

Intermediate hybrids (with quartz and feldspar xenocrysts) have 58.2-63.3 wt% SiO$_2$ and exhibit a scattered trend of elements with increasing SiO$_2$ (Figures 4.1-3). The texturally distinct phenocrystic/xenocrystic rapakivi granitoid hybrids range from 63.5-68.7 wt% SiO$_2$. Hybrids have variable but commonly higher Zr contents than the granites or diorites (Figure 4.2d). On REE primitive mantle (Sun & McDonough 1989) normalized diagrams the hybrids tend to have compositions within the granite to diorite
Figure 4.6  Primitive mantle normalized multi-element and REE pattern of Main Phase granites and microgranites, and fractionated granites of the Sybella Batholith.
Figure 4.7  Primitive mantle normalized (Sun and McDonough 1989) multi-element and REE pattern of Mafic Hybrid Complex, Main Phase granites and hybrids (Rapakivi granitoid hybrids), Sybella Batholith.
range, although hybrids with higher REE contents occur (Figure 4.7). On multi-element diagrams, the more granitic hybrids generally have higher U, and lower Ti contents than the mafic hybrids (Figure 4.7). The scatter is attributed to mixing mafic magma (of a range of compositions – possibly fractionated earlier mafic hybrids) with various proportions of the Main Phase granite magma, which produced compositions intermediate to both end members. Some of the hybrids may have also fractionated at depth and then emplaced as discrete intrusions.

MICROGRANITE HYBRIDS

Field evidence indicates that the hybrids formed at the contacts between the microgranite and Mosses Tank Dolerite (at Kitty Plains) as a result of in situ mixing during emplacement. The analyzed microgranite hybrids (as determined from field and petrographic evidence) plot intermediate to the microgranite and mafic rocks. These hybrids have distinctively different geochemical trends to the rapakivi granitoid hybrids with lower Nb, Rb, Pb and Y (Figure 4.2). On REE and multi-element normalized diagrams the hybrids have compositions generally within the microgranite to dolerite range (Figure 4.8). However, the microgranite hybrids have higher Ba, Hf and Zr than either the microgranite or dolerites.

4.2.5 Country Rocks

The mafic units of the country rocks (May Downs Gneiss and Eastern Creek Volcanics) can be broken up into two geochemically distinct groups: Group 1 includes amphibolites and meta-gabbro rocks; Group 2 consists primarily of mafic gneisses. A mafic intrusion within the Eastern Creek Volcanics and a sample of fine-grained granite from within the May Downs Gneiss was included in the dataset. The latter granite is distinctive in being strongly peraluminous, compared to the weakly peraluminous microgranites of Kitty Plains region. The composition of the mafic country rocks lies within the compositional range for the mafic units from the Sybella Batholith, with the exception of Sr, which is lower. The major element content of Group 1 tends to be with higher TiO₂, Fe₂O₃⁷ and P₂O₅ and lower Al₂O₃ than Group 2 (Figure 4.1). The samples from Group 1 generally have lower REE, Y, Zr, Hf and U than Group 2 (Figure 4.2, 4.9). With increasing SiO₂ the mafic rocks from both groups have constant Y values. The mafic intrusive has a very low REE content with a flat trend. Some of Group 2
Figure 4.8  Primitive mantle normalized multi-element and REE pattern of Mosses Tank Dolerites, microgranites and hybrids, Sybella Batholith.
Figure 4.9  Primitive mantle normalized multi-element and REE pattern of country rocks to the Sybella Batholith. Including: Eastern Creek Volcanics (amphibolite), mafic gneiss, mafic intrusive and fine-grained granite (within May Downs Gneiss). Symbols as for Figure 4.1.
mafic gneisses have compositions intermediate to the fine-grained peraluminous granite (sample JG 96/200) from the May Downs Gneiss. This indicates that some of the felsic melt in the (migmatitic) mafic gneiss may be derived from an external source similar to the May Downs granite intrusion.

4.3 MAGMA MIXING MODELS

There is abundant petrographic and physical evidence of magma mixing between mafic and felsic magmas within the Sybella Batholith (see Chapters 2 and 3). The majority of the elements analyzed from intermediate hybrids (granodiorites) and rapakivi granitoids lie between the geochemical compositions of the Main Phase granite (or microgranite) and the mafic units, which forms a linear trend. This strongly supports the hypothesis that these units have arisen from the interaction between two major bodies. Given the field relationships, the most obvious interpretation to account for the geochemical variation is that magma mixing has taken place. Where magma mixing has been inferred, the mafic (and in some instance the felsic) end-member may not have been identified precisely. The end-member may have been largely consumed in a mixing event or is not exposed. The mafic end-member can be best characterized where there is sufficient variety of “pure”, uncontaminated mafic magmas, showing a trend of chemical evolution (fractionation) that is distinct from the trends interpreted to be the result of mafic-felsic mixing. The mafic end-member that mixed with the microgranites was identified as the Mosses Tank Dolerites within the Kitty Plain region. However, the mafic end-member (or members) that mixed with the Main Phase granites was not determined.

Perfect magma mixing between varying proportions of parental magmas results in linear trends on binary geochemical plots (D'Lemos 1996), and is readily evaluated by statistical parameters. The correlation is a function of the degree to which the perfect mixing has occurred, the homogeneity of each parent magma, the degree to which later modification or mixing has occurred, and the accuracy of the data. A mixing interpretation is supported from the chemical data, with most elements of the hybrid rock lying on straight mixing lines from the mafic to felsic end members for both the
rapakivi granitoid (main phase) and microgranite hybrids (from selected rocks with known associations; Figure 4.10). Mixing within the two suites of hybrids has also been evaluated using Pearson's correlation (Tables 4.2-3). The data set does not give correlation coefficients of +1 or –1 demonstrating that the geochemical variation cannot be explained in terms of perfect mixing behavior. However, the majority of major elements (with the exception of Na$_2$O and Al$_2$O$_3$) do exhibit near linear trends and correlation coefficients to from ± 0.93 to 1, consistent with non-perfect mixing. The absence of perfect correlation could be due to a number of factors, particularly parent inhomogeneity and lack of identified mafic parent. Many of the trace elements (Rb, Sr, Pb, Sc, V, Zn, and Ga) exhibit significant correlation (> ±0.85) with both major and other trace elements. Ba, Th, U, Zr and rare earth elements show a poor correlation (< ±0.85) with other elements. F can be correlated with most major and trace elements in the microgranite suite (Table 4.3), but not in the rapakivi granitoid hybrid suite (Table 4.2). The REE had significant correlation with each other, indicating a near linear relationship between the mafic and felsic end members that is consistent with mixing.

In view of the variable abundance of alkali-feldspar ± plagioclase phenocrysts, the scatter of K$_2$O, Ba and Sr may be ascribed to variable accumulation of feldspars. Scatter of Rb concentrations may result from accumulation of biotite. The poor correlation of some of the trace elements may be due to the low abundances in which they occur, which may give rise to analytical uncertainties (including detection limits and sample heterogeneity). Mixing involving a heterogeneous parent would not necessarily produce linear geochemical trends. Some trace elements are strongly partitioned into accessory phases such as Zr into zircon, and light rare-earth elements into apatite and titanite (Hanson 1978). If such phases formed early in either or both parent magmas any small unevenness in their distribution in the mixed product would result in inhomogeneity on the trace elements in which they concentrate. Fractionation of a mixed magma at depth, such as hypothesized for the rapakivi granitoid hybrids, would also contribute to non-linear relationships.

Magma mixing should result in hyperbolic composition variation in ratio-ratio plots with linear variation in companion plots (Langmuir et al. 1978). Figure 4.11 shows two such ratio-ratio plots and their associated companion plots. The ratio-ratio trends
Figure 4.10  Variation of major and selected trace element contents of rocks with mixing relationships determined from field evidence, from the Sybella Batholith. Trend lines indicate the linearity of some major and trace elements.  (a) 12 samples of Main Phase granite, rapakivi hybrid and hybrid diorite; (b) Microgranite, hybrid and Mosses Tank dolerite (Samples 17 dolerite, 18 dolerite, 27 hybrid, 129 hybrid and 131 microgranite).
Table 4.2 Pearson's correlation coefficient for 12 samples of main phase granites and hybrids (Samples: 177 hybrid dolerite, 183 Granite, 217 hybrid granite, 219 Granite, 307 Mafic Kf hybrid, 312 hybrid, 316 granite, 350 Granite, 351 Granite, 502 Phenocrystic hybrid, 506 hybrid enclave, IS-116 granite). Inset shows correlation of REE.
Table 4.3 Pearsons correlation coefficient for major elements and selected trace elements for 5 samples of microgranite, hybrids and dolerites
(Samples: 17 magnetic dolerite, 18 non magnetic dolerite, 27 hybrid, 129 hybrid, 131 microgranite).
Figure 4.11  Ratio-ratio diagrams and associated companion plots (Langmuir et al. 1978) for the Sybella Batholith: Rb/Sr-Ba/K2O and Rb/Nb-Sc/K2O. Symbols as in Figure 4.1.
(Figure 4.11a and d) are curvilinear and the companion plots are generally linear. The compositional mixing trends for both the Main Phase and microgranite suites with the mafic rocks are different as they occur displaced from each other (Figures 4.11b and 11g). The poor correlation in the mafic units in Figure 4.11c resulted from relatively increased Sr that may be due to cumulate K-feldspar. Figure 4.11f exhibits poor correlation of Sc and Nb, which resulted from variable abundances of mafic silicates and titanite in the least evolved members of the diorites and dolerites. Some combination of mixing and fractional crystallization is required to explain the derivation of the mafic units from a single parental magma.

While magma mixing could account for the majority of field relationships and for petrographic features, the absence of well-defined, linear mixing trends for all elements indicates that perfect mixing did not occur.

4.4 ISOTOPE GEOCHEMISTRY

Nine samples from the Main Phase granite, dolerite, mafic and intermediate hybrids and microgranite were selected for Nd-Sm-Rb-Sr-Pb-U analysis (Table 4.4), and are considered to be representative. The effects of isotopic homogenization related to high-grade metamorphism are notorious, and the disturbance of Rb-Sr systematics was noted in a number of samples. The main objective of this study was to determine the nature of the source region and the mechanisms responsible for the formation of the major intrusive types.

Radiogenic isotopes were analyzed by the University of Adelaide and analytical techniques have been detailed elsewhere (Foden et al. 1995), thus are only summarized here. Following extraction procedures, the Sr-isotopic composition was determined by isotopic dilution and isotopic ratios measured by thermal ionization mass spectrometry on a Finnigan MAT 261 mass spectrometer at the University of Adelaide. Sr- and Nd-isotope analysis followed HF/HNO₃ digestion in high-pressure bombs. Sm- and Nd-concentrations were determined by isotope dissolution after aliquoting and spiking.
4.4.1 The Rb-Sr System

No Rb-Sr isochron was obtained for the samples analyzed as part of this study from the Sybella batholith. For an isochron to be calculated samples from a cogenetic suite (formed from the same parental material) would have been required. The samples analyzed in this study are from the mafic units, Main Phase granite, microgranite and hybrids, and therefore are not suitable for isochron analysis due to their different origin. In addition, the extremely low initial $^{87}\text{Sr}/^{86}\text{Sr}$ in the granites (one of which was strongly deformed) indicates that the isotopic system has been disturbed either by the influx of fluids or by a later thermal event.

Additional Rb-Sr isotope data of the Sybella Batholith main phase granite and microgranite from previous studies (Farquharson & Richards 1970; Farquharson & Wilson 1971) is listed in Table 4.4 and depicted in Figure 4.12ab is plotted with analysis from this study. An isochron was determined from the Main Phase granite samples (as they are a cogenetic suite) resulting in an initial $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7012. The mafic units have calculated initial $^{87}\text{Sr}/^{86}\text{Sr}$ ($T=1670$) of 0.7026 to 0.7034; hybrid units have a higher 0.7053 to 0.7058; a hybrid from Kitty Plains 0.7078; the Main Phase granite 0.6961 to 0.6982 and microgranite 0.6366. The apparent trend in Figure 4.12a results from the mixing of two isotopically distinct reservoirs (end member compositions) that lies along a mixing line. The mafic units have low initial $^{87}\text{Sr}/^{86}\text{Sr}$ values suggestive of a source with relatively low concentrations of radiogenic $^{87}\text{Sr}$ (e.g. mantle) (Figure 12c). The hybrid rocks have higher initial $^{87}\text{Sr}/^{86}\text{Sr}$ values which may indicate a crustal source for at least a proportion of the magma.

4.4.2 Pb isotopes

No U-Pb isochron was obtained for the Sybella Batholith, as they were not from a cogenetic suite. Mafic units have uranium levels below detection, and initial isotope ratios were unable to be calculated. The microgranite has lower calculated initial $^{208}\text{Pb}/^{204}\text{Pb}$ (28.5976) than the Main Phase granite and lies off the linear trend for $^{207}\text{Pb}/^{204}\text{Pb}$ vs $^{206}\text{Pb}/^{204}\text{Pb}$ (Figure 4.13). This disparity may be due to alteration or difference in source composition. The linear trend in the Pb isotopic variation with eNd and SiO$_2$ (Figure 4.13) may be due to mixing (Langmuir et al. 1978) with the hybrids (intermediate in isotopic compositions relative to the mafic and felsic end-members).
4.4.3 Sm-Nd results

Table 4.4 gives new Sm-Nd data for samples of the Sybella Batholith in addition to data from the country rocks from Gunter (1996) that were used in this geochemical study. The intrusions contain between 5.95 and 13.47 ppm Sm, and 27.07 and 75.55 ppm Nd. $\varepsilon$Nd (1670 Ma) values between –6.15 and –2.24 were determined, and from these values, model source ages between 2300 and 2587 Ma ($T_2$) were calculated. The Group 2 mafic gneisses from the country rocks have lower Sm and Nd contents than the Group 1 amphibolites and $\varepsilon$Nd (1860 Ma) values between –0.72 and 0.54 were determined. The mafic intrusive and the felsic intrusive within the country rocks had $\varepsilon$Nd (1860 Ma) values of 2.06 and –3.12 respectively. The $T_2$ model source age was calculated using the method described in Page & Sun (1998) and summarized in Appendix VIII, assuming the mantle depletion commenced at 4560 Ma and was followed by a linear increase to a present day depleted mantle value of +10. $T_2$ was the calculated age that defined the time when the source rocks were separated from the mantle. This age was typically not the age of magmatism, particularly for crustally derived intrusions, where the source was likely to be composed of components that were separated from the mantle at different periods throughout crustal evolution. Thus, $T_2$ represents an average crustal residence time of the source rocks.

The Nd isotopic compositions of the different phases of the Sybella Batholith are shown in an $\varepsilon$Nd versus age diagram in Figure 4.14. The Main Phase granite of the Sybella Batholith contains (64.32 ppm) Nd and (11.40 ppm) Sm, and has a $\varepsilon$Nd of –3.86 (Table 4.5). A $T_2$ model source age of 2419 Ma was calculated using ~ 1670 Ma. The Nd and Sm contents of the main phase hybrids have a positive correlation with SiO$_2$, which is not consistent with fractionation (fractionation of amphibole would partition Nd and Sm from the melt), but rather mixing between the mafic and felsic end member magmas. No mafic end-member was analysed, however a mafic enclave from within the Main Phase granite (with similar geochemical properties to the dolerites from the Mafic Hybrid Complex) had the lowest $\varepsilon$Nd of –6.15 and a T2 model source age of 2587 Ma. The older model source age for the mafic rocks indicated that the melt from the mantle source region (dolerite) was probably contaminated with radiogenically older crustal material (Archaen crust), and that the isotopic signature of this component overprinted that of the mantle. This is also consistent with the enriched LREE, K and Rb contents.
<table>
<thead>
<tr>
<th>Rock Type</th>
<th>Sample</th>
<th>Nd (ppm)</th>
<th>Sm (ppm)</th>
<th>m(^{147})Sm/(^{144})Nd</th>
<th>m(^{143})Nd/(^{144})Nd</th>
<th>i(^{143})Nd/(^{144})Nd</th>
<th>TDM (Ma)</th>
<th>cNd (1860 Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dolerite enclave</td>
<td>322</td>
<td>43,060</td>
<td>8,221</td>
<td>0.1152</td>
<td>0.511429</td>
<td>0.510164</td>
<td>-6.1510</td>
<td>2587</td>
</tr>
<tr>
<td>Hybrid dolerite</td>
<td>177</td>
<td>75,550</td>
<td>13,471</td>
<td>0.1079</td>
<td>0.511391</td>
<td>0.510207</td>
<td>-5.3250</td>
<td>2526</td>
</tr>
<tr>
<td>Mafic Kf hybrid</td>
<td>307</td>
<td>71,097</td>
<td>12,720</td>
<td>0.1082</td>
<td>0.511458</td>
<td>0.510270</td>
<td>-4.0770</td>
<td>2435</td>
</tr>
<tr>
<td>Rapakivi hybrid</td>
<td>314</td>
<td>57,500</td>
<td>10,250</td>
<td>0.1078</td>
<td>0.511458</td>
<td>0.510274</td>
<td>-3.9910</td>
<td>2428</td>
</tr>
<tr>
<td>Main phase granite</td>
<td>351</td>
<td>64,329</td>
<td>11,395</td>
<td>0.1072</td>
<td>0.511458</td>
<td>0.510281</td>
<td>-3.8619</td>
<td>2419</td>
</tr>
<tr>
<td>Mag Dolerite</td>
<td>18</td>
<td>27,074</td>
<td>5,949</td>
<td>0.1329</td>
<td>0.511281</td>
<td>0.510361</td>
<td>-2.2796</td>
<td>2307</td>
</tr>
<tr>
<td>Hybrid</td>
<td>27</td>
<td>33,653</td>
<td>7,275</td>
<td>0.1308</td>
<td>0.511774</td>
<td>0.510338</td>
<td>-2.7486</td>
<td>2337</td>
</tr>
<tr>
<td>Hybrid</td>
<td>262</td>
<td>40,138</td>
<td>8,065</td>
<td>0.1221</td>
<td>0.511663</td>
<td>0.510364</td>
<td>-3.0514</td>
<td>2360</td>
</tr>
<tr>
<td>Microgranite</td>
<td>131</td>
<td>41,457</td>
<td>6,955</td>
<td>0.1015</td>
<td>0.511478</td>
<td>0.510364</td>
<td>-2.2439</td>
<td>2300</td>
</tr>
</tbody>
</table>

* Mafic intrusive JG-96/1A 8.81 2.38 0.16351 0.512337 2.06 2460
* Mafic Gneiss JG-96/40B 27.04 6.17 0.13790 0.511946 0.54 2412
* Mafic Gneiss JG-96/111 20.78 5.02 0.14611 0.511982 -0.72 2622
* Amphibolite JG-96/1B 37.52 9.66 0.14611 0.512110 -0.22 2430
* Amphibolite JG-96/26B 48.45 10.72 0.13373 0.511889 0.42 2394
* Amphibolite JG-96/97A 38.31 9.39 0.14825 0.512068 0.45 2509
* Microgranite JG-96/200 145.85 24.16 0.10013 0.511298 -3.10 2475

Table 4.4 Isotope analysis for Sybella Batholith. Calculations based on formulae Appendix VIII.
Figure 4.12  Measured and calculated initial $^{87}\text{Sr}/^{86}\text{Sr}$ values and trace element variation and relationships between initial $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ (as eNd) within the Sybella Batholith. Additional isotopic data from Farquharson & Wilson (1971) and Farquharson & Richards (1970).
Figure 4.13  Measured and calculated initial \(^{207}\text{Pb}^{206}\text{Pb}\) values and trace element variation and relationships between initial \(^{206}\text{Pb}^{204}\text{Pb}\) and \(^{143}\text{Nd}^{144}\text{Nd}\) (as eNd) and isotopic variation with SiO\(_2\) within the Sybella Batholith.
Figure 4.14  Isotopic harker diagrams of the Sybella Batholith. Initial $^{143}$Nd/$^{144}$Nd (as eNd) at the emplacement age of the Sybella Batholith for intrusive rocks, and at basement age inferred for country rocks. Symbols as for Figure 4.12.
of the intrusion. Contamination in the source region and contamination during transport through the continental crust can be difficult using isotope correlation diagrams alone. Xenocrystic zircons in an amygdaloidal basalt of the Eastern Creek Volcanics reveal the presence of older inherited zircon components of variable ages in two groupings at 2100-1800 Ma and 2700-2500 Ma (Page 1988). These are interpreted by Page (1988) as being relics of the magmatic source and/or from passage through older underlying crust. This is also evidence for the involvement of Archaen crust in the Proterozoic magmatism.

The microgranite contains 41.46 ppm Nd and 6.96 ppm Sm, has a $\varepsilon_{Nd}$ of $-2.24$ and a $T_2$ model source age of 2300 Ma. These values are outside of the range determined for the Main Phase granites and are likely to represent different sources. The sample of Mosses Tank Dolerite has a $\varepsilon_{Nd}$ of $-2.27$ and a $T_2$ model source age of 2303 Ma. The older model source age for the mafic rocks indicates that the dolerite was probably contaminated with radiogenically older crustal material. A Sm-Nd measurement on a sample of B-quartz phase has an $\varepsilon_{Nd}$ value of $-3.1$ and a depleted model age of 2336 Ma (Page & Sun 1998).

### 4.5 GEOCHEMICAL INTERPRETATION

Successful models for the origin and evolution of the Sybella Batholith must explain the A-type characteristics of the Suite (Fe-rich, low compatible element concentrations), the low $fO_2$ and high $T$ conditions of emplacement, and the compositional variation.

Geochemical trends within the mafic units are attributed to a combination of fractionation and mixing, whereas mixing with the granite phases produced a suite of intermediate hybrids. Fractionation within the felsic units was a minor process as there is little geochemical variation within the Main Phase granite, although the relatively small volume of porphyritic granite to aplite (texturally related to the Main Phase granite) that intruded the Main Phase granite and mafic units indicates that it did occur.
The broad geochemical trends within the mafic units are consistent with a combination of fractional crystallization of observed mineral phases and mixing. The negative correlation between CaO, Fe₂O₃T, MgO, MnO and SiO₂ is consistent with the fractional crystallization of pyroxene, and to a lesser extent amphibole (Figure 4.1). The decrease in Fe₂O₃T (and V) may also relate to the fractionation of magnetite. The slightly negative correlation between Ba and Rb for the Main Phase granites observed in Figure 4.3g possibly results from minor fractional crystallization of K-feldspar. The lack of linear trend in K/Rb with increasing Rb (Figure 4.3h) indicates that extensive K-feldspar fractionation did not occur. It is also noted that most samples exhibit K/Rb ratios greater than 160, which indicates control by crystal-melt equilibria (Shaw 1968), and suggests that magmatic processes controlled the evolution of this complex with minimal involvement of post-magmatic rock-fluid interactions (which typically produce K/Rb ratios of less than 160; Abdel-Rahman & El-Kibbi 2001). The microgranite (DC-25/1-120) and hybrid (IS-19.2-108) have the lowest K/Rb ratios indicating that post-magmatic albitionization may have played a role.

The granite and dolerite phases within the Sybella Batholith have REE signatures that are not consistent with the two phases being derived by the fractionation of minerals from the same parent melt. The Main Phase granite and hybrids contain slightly higher HREE than the dolerites (Figure 4.7), which is inconsistent with typical pyroxene (or hornblende) fractionation. Pyroxene or hornblende fractionation generally lowers the HREE content of granitoids with increasing fractionation (Rollinson 1993), which is the reverse of the Sybella Batholith trend. Although the trend of undepleted Eu in the dolerites to depleted Eu in the granites could have been produced by the preferential partitioning of Eu (and Sr) into plagioclase (Rollinson 1993) during crystallization, two different sources (a mafic undepleted source and a granitic depleted source) that mixed to produce a range of depleted magmas is considered to be more probable. The felsic granitoids (Main Phase and microgranite phase) are enriched in LREE relative to the mafic rocks and have high HREE (with little to no change in HREE with apparent evolution of the granites), which could indicate that relatively higher concentrations of pyroxene were present in the source region to produce a relative enrichment in LREE (La) with respect to HREE (Lu). These geochemical characteristics, along with petrographic and field evidence, indicate that the rapakivi granitoids are probably a
product of magma mixing or an interaction of mafic magmas and feldspar-rich crystal mush. REE patterns for the Mafic Hybrid Complex, Main Phase granites and their hybrids are generally similar in shape.

The high-K nature (Figure 4.1) of the granite intrusions suggests that their source contained a substantial concentration of K-bearing silicates (e.g. biotite and K-feldspar). The effect on the K_2O content of the starting material was illustrated by Roberts & Clemens (1993). To produce a high-K melt the most suitable source material are calc-alkaline and this implies significant volumes of calc-alkaline source rocks at depth. Partial melting of these rocks produces melt of high-K character and therefore does not need metasomatised source regions or any other chemical or physical process.

The Sr-depleted, Y-undepleted nature (Figure 4.5-8) of the complex indicates that plagioclase was present and garnet was absent from the source region during melting (Wyborn et al. 1987), and that melting probably occurred at <8-10 kbar (Wyllie & Wolf 1993), which would inhibit the formation of garnet in amphibole- or pyroxene-bearing rocks. The high melt temperatures of ~ 900°C and the H_2O undersaturated conditions of crystallization (Holtz et al. 1996) for the Main Phase granites are consistent with melts that formed via fluid absent partial melting which was induced by the emplacement of mantle-derived material into the crust (Ramo & Haapala 1995; McCarthy & Patino Douce 1997). Dehydration melting of calcalkaline granitoids can give rise to A-type granitic melts, but this is only true if melting takes place in shallow crust. Melting experiments of tonalite and granodiorite at 950°C indicate that melting at low pressures (5-7 kbar) results in more definite A-type characteristics than at 8 kbar (Patino Douce 1997), and that high temperatures >900°C are needed to obtain significant melt fraction from H_2O poor calc-alkaline source material. High temperature melting also leads to the enrichment of HFSE due to the increased solubility of accessory minerals. The low water content of the protolith explains the H_2O poor nature of the A-type magma (Patino Douce 1997). Therefore, the water to produce melting is interpreted to have formed via a combination of biotite (Reaction 1) and amphibole (Reaction 2a and b) dehydration reactions (Wyllie & Wolf 1993).

\[
\text{Biotite + quartz + plagioclase} \pm \text{K-feldspar} \rightarrow \text{melt} + \text{K-feldspar} + \text{pyroxene} (1) \\
\text{Hornblende + plagioclase} \rightarrow \text{melt} + \text{pyroxene} (2a) \\
\text{Hornblende} \rightarrow \text{melt} + \text{pyroxene} + \text{plagioclase} (2b)
\]

4-33
The lower concentration of Zr, Y, and distinctive REE patterns in the microgranites compared with the Main Phase granite, and the difference in the T₂ model source age preclude a direct petrogenetic link with the Main Phase granites. The microgranites may be derived from a different source or from successive melting of the same source. Some of the geochemical relationships between the microgranite and the Main Phase granites may be accounted for by K-feldspar and zircon fractionation. However, the low Zr, REE and Y values in the microgranites may result because the granite is a partial melt from a different source, in which the accessory minerals were left in the residue and therefore are not genetically related to the Main Phase of the Sybella Batholith. The low abundances of Sr, Ba, Zr and the REE (especially Eu) in the microgranites suggest a source with residual plagioclase, alkali-feldspar, zircon and allanite. The re-melting or batch melting of the same source would invoke that the microgranite was derived from the residual material after the production and egress of the potassic magmas. This scenario may in part be highlighted by the higher εNd and lower model source ages, which indicates a more juvenile component in the source region.

The negative εNd values of the Main Phase granite, microgranite and mafic units are an indicator of crustal involvement in magma processes. These intrusions were derived from a source region with an average Palaeoproterozoic residence age between 2587 to 2300 Ma. The older model source age for the mafic rocks indicates that the melt from the mantle source region was probably contaminated with radiogenically older crustal material, and that the isotopic signature of this component overprinted that of the mantle. Contamination in the source region or contamination during transport through the continental crust was unable to be determined.

Therefore, with the exception of the K-feldspar porphyritic granites and Main Phase granites, none of the granitic units share a liquid line of descent with any other granitic unit. This suggests that the Sybella Batholith contains evidence for emplacement of at least three distinct types of magma; the Main Phase granite (and porphyritic granite), the microgranite and the mafic magmas, which indicates a combination of compositionally heterogeneous crustal source and mantle source. The mafic magmas show strong Fe-enrichment which indicates fractionation occurred within the crust, and that they were
not directly sourced from the mantle. These enriched tholeiites have a Sr-depleted and Y-undepleted pattern that is similar to the granites of the Sybella Batholith, which indicates interaction with the granites or similar partial melts at depth.

4.6 DISCUSSION

An outline of the origin and evolution of the magmatic components of the Sybella Batholith is schematically illustrated in Figure 4.15.

Production of large volumes of high-T granitic magma in the continental crust is commonly attributed to emplacement of mafic material that triggers partial melting in the lower to middle crust (e.g. Huppert & Sparks 1988; McCarthy & Patino Douce 1997). In the Sybella Batholith, gabbroic to dioritic rocks represent a significant component, and points to mantle involvement in magma genesis. The shallow origin of A-type granites is in turn a consequence of their setting in a noncompressive tectonic regime, where the crust tends to thin (Patino Douce 1997). Basaltic magmas will intrude to a level where there is an effective density barrier, and then spread out as horizontal intrusions. The focus of intrusion might be the Moho or the middle crust at the transition from dense lower crust to comparatively lower density upper crust. In the initial stage, the crust is cold enough to allow some basalt to penetrate to the surface, with chilling of dyke and sill margins (Figure 4.15.a). Each individual input of basalt increases the heat content of the crust. With the crust becoming hotter, the amount of crustal melting associated with each input increases with time (Figure 15.b). Huppert & Sparks (1988) modeled the thermal effects of intrusion of basaltic magmas into continental crust, with calculated temperatures between 900-950 °C if there is thermal convection in the mafic magma. The model proposed by Roberts & Clemens (1993) involved fluid absent partial melting of metaluminous protoliths (metamorphosed igneous rocks in the lower crust) that were heated by underplated or intraplated mafic magma (e.g. Clemens 1990). The potassic incompatible- and radiogenic element-rich character of the Sybella Batholith suggests that the source material was fertile in these elements, and that these intrusions were likely to be the initial melt products of a virgin
source. The high temperature of the magma is evident from the lack of restite (Wyborn et al. 1988).

With increasing temperature at the focus of the intrusion builds up and as the efficiency of trapping basalt in hot crust increased. The crustal rocks above the mafic sill came to have a lower fusion temperature than the magma, which led to the melting of the roof rock. Extensive partial melting of the crust was initiated and the system evolved predominantly to silicic magmas. A stable layer of (lower density) silicic magma formed and grew with time. As the thickness of the silicic magma increased, thermal convection eventually began.

In addition to crustal melting, some mantle-derived mafic magma mixed with crustal magmas to form mafic hybrids. Hybridization is possible between dry silicic high-temperature magmas and mafic magmas in the upper crust when their viscosities are similar (Frost & Mahood 1987). Magma chambers with basic and silicic magmas, and zones of mixing and mingling can form deep in the crust where mantle-derived basic magmas cause partial melting of the more silicic crust (Huppert & Sparks 1988). Repeated injections of basic magma into the magma chamber can produce zones of mingling and mixing and result in hybrid magmas. Intrusion of magmas from such chambers can produce rapakivi granites, dolerite and gabbroic rocks, and hybrid complexes. Several facts suggest that the hybridization took place at a deeper level before the magmas were emplaced to form the batholith. First, the relatively homogenous hybrid rocks are widespread although they are volumetrically minor components of the batholith. Second, there is a lack of mafic rocks within the Mafic Hybrid Complex that show no mixing with felsic magmas, which indicates that the mafic magma had generally hybridized to some extent before emplacement. The ability of two contrasting magmas to mix depends on their physical properties, especially viscosity, after they have come to thermal equilibrium (Sparks & Marshall 1986). Complete mixing can occur when the two magmas behave as liquids at the same temperature. This requires a substantial proportion of basic magma to bring enough heat into the system and time for chemical diffusion. The mafic hybrids intruded into the May Downs Gneiss and Eastern Creek Volcanics as variably hybridized batches. Emplacement of intermediate rapakivi hybrids into the mafic hybrids occurred when the
mafic hybrids were still molten, leading to further interaction (mingling and mixing) at emplacement level.

Magma mixing has been a major mechanism in generating hybrid rocks of the Sybella Batholith, but other mechanisms have also contributed to its magmatic evolution. Fractional crystallization (at depth) is possibly a significant contributor to the variation in the mafic rocks, however the extent of fractionation in the felsic rocks appears limited. The basaltic layer, which cooled and crystallized as it lost heat, eventually became so crystal-rich that it effectively was a solid and convection stopped. However, the overlying silicic layer continued to convect and generated further melt at its roof. Finally, the silicic layer itself crystallized and convection ceased (c.f. Huppert & Sparks 1988).

Magmas are generally thought to result from crystallization of almost crystal-free magmas, emplaced with no or limited fractionation during ascent. The above model allows for extensive crystallization in the source region. Large volumes of phenocryst-rich magmas could be generated in the source region over a short period. Field and petrographic characteristics indicate that Main Phase granites within the Sybella Batholith were emplaced as a phenocryst-rich magma. The idea that melting occurs in the source, and that crystallization only occurs after the magma has ascended to a shallow level in a cooling environment, is not valid in this particular case (Huppert & Sparks 1988). However, this model does not preclude the emplacement and subsequent crystallization of crystal-poor magmas at shallow depth.

The occurrence of contemporaneous mafic and felsic magmas and Fe-enrichment within the Sybella Batholith, is a common feature in other ‘A-type’ rapakivi granites (Ramo & Haapala 1995). The Main Phase granite intruded into the Mafic Hybrid Complex, and contains mafic and hybrid enclaves (molten), and xenoliths (solid). Porphyritic granite and aplite then intruded the complex. The limited development of pegmatite of similar age suggests that conditions remained largely anhydrous throughout crystallization of this suite. Similarly, in the northern region of Kitty Plains, the microgranite intruded into a complex dolerite body (Mosses Tank Dolerites) and local hybridization and
brecciation occurred along the contacts. The lower temperature determined for the microgranite probably inhibited the production of homogenous hybrids.

The global distribution of Proterozoic intrusions with K-rich and ‘A-type’ geochemical characteristics and late-Archaean to Palaeoproterozoic model source ages (Ramo & Haapala 1995), suggests the occurrence of a widespread magmatic event. The Sybella Batholith is interpreted to be part of this globally distributed group of dominantly Proterozoic (mainly 1.0 to 1.7 Ga), high-T, K-rich intrusions with ‘A-type’ geochemical characteristics, which were produced by temperature-induced melting via the injection of mafic material into the lower crust (Ramo & Haapala 1995).
Chapter 4  

Geochemistry

Figure 4.15  Schematic development of the Sybella Batholith.
(A) Subcrustal lithospheric extension with inset representing the location of Sybella Batholith in the western Fold Belt, Mount Isa Inlier. Modified from Betts et al. (1998) and MacCready et al. (1998).

MODEL

(B) Initial basalt/plug formation.  (C) Continued influx of mafic magma into lower crust resulted in partial melting and production of felsic magmas. The mafic magma interacted with the felsic melt resulting in a hybrid magma. This magma was emplaced into the Cover Sequences as Mafic Hybrid Complex. (D) With continued influx of mafic magma into the lower crust, a stable felsic layer developed. When it was emplaced the felsic intrusion (main phase granite) mingled and mixed with the mafic hybrid complex. (E) Shear zone developed during the intrusion of crystal-rich mush of the Main phase granite, as a result of continued extension. (F) Crystal liquid separation of felsic melt (or melting of a different source) formed the microgranite magma. Intrusion resulted in local mingling and brecciation with the earlier mafic intrusions.